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Economic Geology

and the

Bulletin of the Society of Economic Geologists

CONTENTS

Present Status of Our Knowledge Regarding the Hydraulics of Ground Water ..... *O. E. Meinzer and L. K. Wenzel* 915

The Spodumene Pegmatites of North Carolina ..... *Frank L. Hess* 942

The Molybdenite Deposits of the Rencontre East Area, Newfoundland ..... *D. E. White* 967

Geology of the Nighthawk Peninsular Gold Mine ..... *A. Roddick Byers* 996

Discussion and Communications:

Openings Due to Movement Along a Fault ..... *N. D. McKechnie* 1012

A Theory of Mineral Sequence in Hypogene Ore Deposits ..... *L. H. Hart* 1014

Reviews:

Australia, a Study of Warm Environments and Their Effect on British Settlement (Taylor) ..... *W. S. Bayley* 1019

Sedimentary Petrography, 3rd edit. (Milner) ..... *W. S. Bayley* 1019

The Physical Sciences (Cable, Getchell and Kadesch) 1021

Commodities in Industry, 1940 Year Book 1021

Books Received ..... *William E. Benson* 1022

Scientific Notes and News 1024

Index, Volume XXXV 1025

Contents, Volume XXXV iii

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PRESENT STATUS OF OUR KNOWLEDGE REGARDING THE HYDRAULICS OF GROUND WATER.*

O. E. MEINZER AND L. K. WENZEL.

ABSTRACT.

Darcy's law, that the flow of water through porous materials increases directly as the hydraulic gradient, applies to laminar flow, which is characteristic of ground water. The rate of movement of the ground water has been determined at many places; its movement is caused by the difference in head between intake and discharge. The hydraulic gradients are three-dimensional. In stratified formations the water moves chiefly parallel to the stratification but movement through the confining beds is an important factor in the recharge and discharge of artesian aquifers.

Artesian systems possess volume elasticity and hence variations in storage capacity with changes in the internal buoyant force due to changes in head. Presumably the compression, and also the re-expansion, occur largely in the strata of relatively fine grain, which feed into the strata of coarser grain that supply the wells. The phenomenon of volume elasticity has been given mathematical study by C. V. Theis, who applied the theory of the flow of heat, and by C. E. Jacob, who has attacked the problem directly with hydrologic concepts.

Dupuit applied Darcy's law to the hydraulics of wells and developed the widely-known equilibrium-type formula. G. Thiem employed the Dupuit formula for determining permeability of water-bearing formations by using two observation wells. Theis introduced the non-equilibrium formula for determining drawdown at any time and at any place on the cone of depression.

There are four groups of methods for determining the permeability of water-bearing materials: (1) direct laboratory methods, (2) indirect laboratory methods, (3) field velocity methods, and (4) field discharge methods. There are now in use in the United States Geological Survey a number of interrelated field discharge methods, which are expressed by formulas designated as the Thiem formula and associated limiting formula, the gradient formula, the non-equilibrium formula, and the recovery formula.

* Published with the approval of the Director of the Geological Survey, U. S. Department of the Interior.

CONTENTS.

Laminar and turbulent flow 916
Darcy's law 917
Course and rate of movement of ground water 918
Head in relation to movement 919
Permeability of water-bearing materials 921
Storage in relation to movement 922
Hydraulics of wells 926
Dupuit formula 926
Thiem formula 928
Limiting formula 929
Behavior of ground water in the vicinity of discharging wells .. 931
Gradient formula 933
Non-equilibrium formula 935
Recovery formula 938
Extent of cone of depression 938

LAMINAR AND TURBULENT FLOW.

THE water in the interstices of the permeable rocks in the zone of saturation is, as a rule, moving very slowly and very steadily. This steady kind of movement is called laminar flow, also stream-line or viscous flow. In each thread of the laminar movement there is an endless procession of particles of water, following the same path, propelled by the differential pressure that overcomes the friction with the adjacent thread of more slowly moving particles nearer the walls of the openings through which it passes. The whole process is conceived to be perfectly orderly, each particle moving in a path that is predetermined by the irregular shapes of the interstices and the three-dimensional system of hydraulic gradients adjusted to these interstices—urged, ever forward, gently but persistently.

In contrast to laminar flow, is turbulent flow, in which eddies occur and the water particles move in irregular circuitous paths. The flow in rivers and creeks is generally turbulent whereas the flow of ground water is laminar except under unusual conditions, as perhaps in some underground streams in cavernous limestone or where the ground water is entering a heavily-pumped well.[21]

The nature of the two modes of flow was demonstrated by

Osborne Reynolds [5] in a series of experiments on parallel glass tubes of different diameters up to two inches. The velocity at which eddy formation is first noted in a long tube is termed the higher critical velocity. There is also a lower critical velocity, at which the eddies in originally turbulent flow die out. The critical velocity varies directly with the viscosity and hence is higher in cold than in warm water. It varies inversely with the diameter of the tube and directly with the roughness of the walls of the tube.

DARCY'S LAW.

The flow of water was first studied in open channels and pipes, in which turbulent flow occurred, and it was found that the rate of flow varied approximately as the square root of the hydraulic gradient. The flow of water in capillary tubes was first studied by Hagen [1] and Poiseuille,[2] who discovered that the rate of flow through these small tubes varies directly as the hydraulic gradient. Later Darcy [3] verified this law and demonstrated its application to water percolating through interstices of filter sand.

Since the results of Darcy's work were published, there has been much discussion as to whether this law expresses closely the law of flow of water through porous material, whether it is applicable only through a certain range of velocities and hydraulic gradients, and if so, what are the lower and upper limits. Many laboratory investigations have been made on the flow of liquids and gases through permeable materials but the whole subject was left somewhat obscure.[6]

In 1923 the hydrologic laboratory of the Geological Survey was organized and systematic investigation was undertaken, under the direction of the senior author of this paper, of the flow of water through permeable materials under the low gradients that occur naturally in ground water.[11] In tests made in this laboratory on about 2,000 samples of material, from many different water-bearing and non-water-bearing formations, the validity of Darcy's law has been thoroughly established. In 1933, Fishel and the senior author obtained results, for the material tested, showing that Darcy's law holds down to a gradient of 2 or 3 inches to

the mile, and that there are indications that it holds to less than 1 inch to the mile.[15]

It is now clear that the simple law, that the flow increases directly as the hydraulic gradient, applies to the laminar flow, in which no energy is lost in producing eddies. The failure of some of the early investigators to verify Darcy's law was probably due largely to the fact that they performed all or a part of their experiments with gradients so high that the flow through the porous media was of the turbulent type. The upper limit of the application of Darcy's law to ground water is therefore the critical velocity.

The results of Reynolds' work were published in 1883, but the whole subject of the upper limit of Darcy's law was confused about the same time by Daniell, who defined capillary tubes as those in which the flow of water follows Darcy's law and gave the limiting size of such tubes as one-fiftieth of an inch in diameter. It was further confused by Van Hise, who expressed one-fiftieth of an inch as .508 millimeter, thus giving an appearance of accuracy that was fictitious.* The whole subject of the laminar and turbulent flow of water is effectively presented by Tolman (20: 191–200) in his recent text book. Tolman clears up the confusion by pointing out that the idea that there is a limiting size of tube that determines the type of flow is incorrect and that laminar flow takes place in tubes of any size provided the velocity is sufficiently low. On the basis of tests made by J. F. Poland and others, he shows that under normal ground-water gradients, turbulent flow in sands and gravels is virtually non-existent.

It has long been recognized that Darcy's law is analogous to the fundamental law of the conductivity of electricity and heat but that there are certain important differences in application.

COURSE AND RATE OF MOVEMENT OF GROUND WATER.

In general each water particle in the zone of saturation is moving from some point in an intake area where it first reached the water table toward a point where it is to be discharged even-

* See U. S. Geol. Survey Water-Supply Paper 489, p. 26.

tually through a subaerial, sublacustrine, or submarine spring or seep, or by evaporation or absorption by the root of a plant. The path through which the water particle threads its way may be short and simple, never getting far below the water table, or it may be many miles—indeed as much as a few hundred miles—in length, and with a circuitous course, leading perhaps to depths of hundreds or thousands of feet. The time consumed by the water particle in making the trip from source to exit may be relatively brief or it may be many years or many centuries.

The rate of movement of the ground water has been determined at many places—directly by the use of salts or dyes and indirectly from determinations of permeability.[18] In field tests, rates of movement of as much as 420 feet a day have been found whereas the lowest rate observed in the hydrologic laboratory was about one foot in 10 years.[11, 17] The total range in nature is doubtless even greater. However, in the recognized water-bearing formations, from which the wells obtain their water supplies, the natural rate of movement of the ground water is generally not greater than 5 feet a day and not less than 5 feet a year.

HEAD IN RELATION TO MOVEMENT.

As the water moves, energy is expended in overcoming friction and is converted into heat or perhaps into some other forms of energy. The movement is so slow that the energy consumed in overcoming the inertia of the water is relatively negligible. The energy that keeps the water in motion against the internal friction created by its own viscosity is provided by the difference in head between the place of intake and the place of discharge. The path followed by each thread of water particles may lead in varying directions, and in parts of its course it may have an upward trend, but the exit is at a lower level than the intake, and thus potential energy is lost. The difference in head is distributed throughout each thread of water as an hydraulic gradient continuously, but not at a constant rate, all the way from intake to exit.

In the investigation of a given area, the ground-water hydrologist generally makes a contour map of the water table because it throws much light on the questions as to the direction

of movement of the ground water and hence as to the intake and discharge areas. Strictly, however, even a perfect contour map of the water table would show only the horizontal direction of movement of the ground water at the water table. The hydraulic gradients are actually three-dimensional, and the water moves not only along the water table but also to depths below the water table and generally upward again to the water table at some other place. Obviously, where the water is moving downward there is a decrease in head with depth, and where it is moving upward there is an increase in head with depth. Thus, even in a deposit of structureless sand the head at some depth is likely to be different than at the water table—lower if it is in an intake area and higher if it is in a discharge area.

The sedimentary rocks are, however, not structureless but are generally composed of successive strata that differ greatly in permeability. Consequently the ground water moves chiefly along the stratification, through the most permeable strata, and much less freely through the less permeable strata. In applying Darcy's law, however, it is evident that although in movement at right angles to the strata the permeability factor is likely to be very small, the other two factors—hydraulic gradient and cross-sectional area—are likely to be relatively very great in comparison to the gradients and cross-sectional areas in movement parallel to the stratification.

Where the permeable strata lie between relatively impermeable strata, the latter tend to form more or less effective confining beds. If the strata are tilted or deformed, the water may be led from the intake areas of the permeable strata for long distances and to great depths through the conduits thus formed. It is a common fallacy to say that ground water always moves down the dip. In fact it generally moves down the dip for some distance from the intake area, but in its later course it may move down or up according to the deformation of the strata, as, for example, in the Dakota sandstone and in the synclines that give rise to most of the thermal springs in the Appalachian area.

If the loss in head is less than the net descent of the water-bearing strata, the water comes to be under artesian pressure in

the sense that it will rise in wells to some level above the top of the aquifer; if the loss in head is less than the descent of the land surface, the artesian pressure may be sufficient to cause the wells to overflow. Many other structural features besides tilted or deformed strata may also produce artesian conditions, but the general principles are the same.

A piezometric surface of any aquifer is an imaginary surface which everywhere coincides with the head of the water in the aquifer. In areas of artesian flow it is above the land surface. Contour maps of piezometric surfaces are useful in much the same way as contour maps of the water table, in indicating the general direction of movement of the water and hence its approximate source and destination. Like water-table maps, they show only the horizontal direction of the movement, whereas if the confining beds are not entirely impermeable, there is generally upward or downward leakage through them. Thus, if the water above the overlying confining beds is under greater head than the water in the confined aquifer, there will be percolation into the aquifer, but if it is under less head, there will be loss of water from the aquifer by upward escape. In many of the artesian systems the movement of water through the confining beds is an important factor in the recharge and discharge of the aquifer and in the building up or loss of head.

PERMEABILITY OF WATER-BEARING MATERIALS.

The hydraulic permeability of a porous material is its characteristic property of transmitting fluids through its interstices. It is a property analogous to the conductivity of materials for heat and electricity. The degree of this property has been designated by different names and has been expressed by various combinations of units of space and time. It is now rather widely called the coefficient of permeability.

Recently C. V. Theis[16] introduced the convenient term *coefficient of transmissibility,* which is the coefficient of permeability, with a temperature correction, multiplied by the thickness of the saturated part of the aquifer. Thus, the coefficient of permeability denotes a characteristic of the water-bearing material,

whereas the coefficient of transmissibility denotes the analogous characteristic of the aquifer as a whole.

Natural earth minerals that have been tested in the hydrologic laboratory of the Geological Survey have been found to have coefficients of permeability ranging from about 0.0002 to about 90,000; that is, the most permeable material carries water at a rate of about 450,000,000 times that of the least permeable material. However, most water-bearing materials utilized by wells have coefficients that are whole numbers of two or more figures, generally between 10 and 5,000.

As the permeability of water-bearing materials is of fundamental importance in the study of ground water, it is not surprising that much work has been done in developing accurate and feasible methods of determining this property. The different methods may be grouped as (1) direct laboratory methods, (2) indirect laboratory methods, (3) field velocity methods, and (4) field discharge methods.[18] The methods now chiefly used are the direct laboratory methods and the field discharge methods.

There are now in use in the Geological Survey a number of somewhat interrelated field discharge methods or groups of methods which are expressed by formulas designated as (1) the Thiem formula and associated limiting formula, (2) the gradient formula, (3) the non-equilibrium formula, and (4) the recovery formula. These methods are described later in this paper.

STORAGE IN RELATION TO MOVEMENT.

The quantity of water stored in the interstices of the rocks is much less than the quantity in the ocean but many times the quantity that occurs at any time in the lakes, streams, and atmosphere. Investigations have shown that the Carrizo sand in a 60-mile stretch in the winter garden region in Texas, transmits about 24,000,000 gallons a day at an average velocity of only about 50 feet a year. Accordingly each mile of the formation extending from the outcrop toward the coast contains in storage a supply equal to about 24,000,000 gallons a day for a century. It should be noted, however, that most of this stored water is not economically available for use.

The slow movement of water through the permeable rock formations is comparable to the movement of water through a surface lake or reservoir from the mouths of the tributaries to the outlet. As has been previously pointed out by the senior author, some aquifers function chiefly as reservoirs and others chiefly as conduits, but all have some of the properties of both.[13] In all adequate ground-water studies these two functions must be recognized and differentiated.

It is generally considered that, for the most part, the formations or parts of formations that yield water with only moderate contents of dissolved mineral matter form parts of circulating systems in which water of meteoric origin flows from the intakes to the outlets of the formations, ultimately flushing out the salty connate water; and conversely to a great extent the formations or parts of formations that contain salty water have little or no circulation.

It is believed that some of the artesian aquifers of the Atlantic and Gulf Coastal Plain have submarine outcrops that form potential outlets for the artesian water, whereas others pinch out or become impermeable at their seaward edge. In the aquifers that have outlets the fresh artesian water presses against the heavier sea water. If the head of the fresh water is relatively high and the submarine outlet not too far below sea level, the fresh water discharges into the sea. Otherwise the sea water backs up into the aquifer and maintains a static condition, except as the fresh water may escape upward through the confining beds. Flowing and pumped wells reduce the head and may thus cause the sea water to percolate toward them. To some extent the existing conditions may have been inherited from the Pleistocene epoch, when the sea stood at times higher and other times lower than at present. That salty water is definitely in circulation in some places is shown by the occurrence of large salt-water springs.

If there were no intake or recharge, the ground-water systems would gradually run down, somewhat like a clock that is not wound. But the ground-water systems form parts of the "hydrologic cycle." From time to time the rain and snow descend upon the earth and a part of the water finds its way down to the

water table. Whenever and wherever this occurs new water goes into storage in the aquifer, the water table is built up, the head is increased, new energy is supplied, and the movement of the ground water is quickened.

Recharge is never uniformly distributed but occurs chiefly in certain localities that are favorable for intake. Hence, the water table is built up irregularly, its shape is changed, and mounds are developed on it in the principal intake areas. Hence, the whole three-dimensional system of hydraulic gradients becomes disturbed, new lateral components of these gradients are developed, the direction of movement of the ground water is modified, and many of the water particles change their courses and head for new outlets. Whenever the water table is lowered in any discharge area—as at springs, in localties of ground-water evaporation and transpiration, or by the pumping or artesian flow from wells—a series of changes occur which are the converse of those that result from recharge. These irregular changes in head make the hydraulics of ground-water complex.

The adjustments that result from disturbances in the head do not occur instantaneously because they require the transfer of water from the localities of high head to those of low head. Moreover, the ground-water systems are not entirely like clocks, which run down at a uniform rate and come suddenly to a stop, but they approach equilibrium with constantly diminishing speed. This fact has been recognized more or less for a long time in so far as water-table conditions are concerned.

Until rather recently it has been tacitly assumed that artesian systems are rigid and that therefore changes in storage can occur only near the outcrops of the aquifers where a water table exists. Under such conditions the adjustments caused by changes in head would occur quickly, as no transfer of water would be involved except the small amount due to the slight volume elasticity of the water.

In studies of the Dakota artesian system by the senior author in 1923 and 1924, he recognized from several lines of evidence, that the artesian system does not perform like a rigid system but as one having volume elasticity and hence variation in storage

capacity with changes in the internal buoyant force due to artesian head. He found evidence that a large part of the water that had been discharged from the flowing wells was water taken from storage by compression of the water-bearing beds with loss of the artesian head. Consequently, he stated the theory of the compressibility and elasticity of artesian aquifers,[10, 12] which has since been verified by observations in many other areas. The phenomenon of compressibility has been found to be of great practical importance in the study of the perennial yield of artesian aquifers, it is significant with respect to the recovery of petroleum, it is known to be responsible for the subsidence of the land surface in some heavily pumped localities and it will probably be found to be responsible for subsidence in many others. Presumably the compression, and also re-expansion, occur largely in the strata of relatively fine grain that feed into the strata of coarser grain that supply the wells.[19]

It has generally been considered that under artesian conditions there is very little lag in the surrender of the water due to compression. However, it now appears that if the principal compression occurs in fine-grained deposits that have low permeability and may be connected only rather indirectly with the most permeable part of the aquifer, there may be much more lag than has been assumed and the quantities of artesian water removed from storage may be greater than is indicated by the coefficients of storage determined in tests of short duration.

The non-equilibrium method developed by Theis with the aid of C. I. Lubin attacks the problem of the hydraulics of ground water before equilibrium is reached, thereby involving time as a variable.[16, 23] It applies the mathematical theory of heat conduction developed by Fourier and subsequent writers to ground-water hydraulics. As Theis stated, this analogy between ground water and heat has been recognized at least since the work of Slichter, but apparently no successful attempt had been made to introduce the function of time into the mathematics of ground-water hydrology. With this formula it has become possible to use the hydrologic data in computing the long-term effects of pumping from wells or of other occurrences that produce changes

in head, and it has led to new methods of determining permeability and storage coefficients. Recently Jacob developed the non-equilibrium formula directly from hydrologic concepts without recourse to the analogy of heat.[24]

HYDRAULICS OF WELLS.

Because of the importance of the development of water supplies from wells and of the other human activities that affect the ground-water regimen, the subject of the hydraulics of the ground water is of great practical significance. These artificial disturbances, especially the withdrawal of water from wells, upset the approximate equilibrium of nature and create complicated conditions that make the hydraulics of the ground-water complex. However, they also afford field laboratories in which the effects of definitely regulated changes can be observed and measured.

Dupuit Formula.—The setting forth by Darcy[3] in 1856 of the law of ground-water movement made possible for the first time the mathematical treatment of the hydraulics of wells. Dupuit[4] was apparently the first to undertake mathematical treatment. He assumed the existence of a circular island in the precise center of which is situated a well that extends through a permeable water-bearing formation that is homogeneous in character, an aquifer that rests on a horizontal impervious stratum, a water table that extends horizontally beneath the entire island and is contiguous to and level with the surface of a vast open reservoir whose supply is unlimited (Fig. 1). Dupuit assumed further that the well is pumped at a constant rate for such a period that the cone of depression of the water table created by the pumping extends to the periphery of the island. A condition of hydraulic equilibrium then prevails—that is, the slope of the water table everywhere beneath the island is just sufficient to transmit water continuously through the aquifer from the open reservoir to the well at a rate equal to the rate of withdrawal from the well. With equilibrium thus established the flow through each concentric cylindrical section around the well is equal to the flow through every other cylindrical section and is equal also to the discharge of the well.

According to Darcy's law, the flow through each of the concentric cylindrical sections of water-bearing material is equal to PiA, where P is the coefficient of permeability, i is the hydraulic gradient and A is the cross-sectional area of the cylinder. If x is taken to be the radius of any cylinder and y to be the saturated thickness of material at the distance x from the well, the hydraulic gradient i, is then equal to the rate of change of the coordinates —that is, dy/dx, and the cross-sectional area of the cylinder is

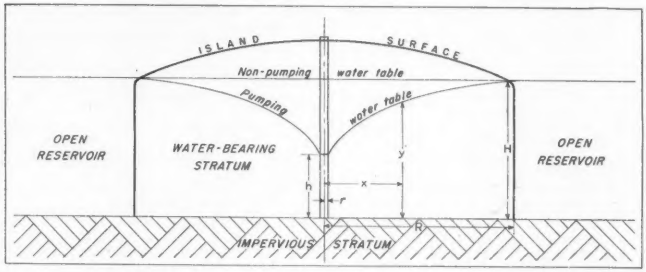

FIG. 1. Ideal conditions assumed for the development of the Dupuit formula for water-table conditions.

equal to $2\pi xy$. The following formula can then be written to express the relation of the rate of discharge of the well (Q) to the shape of the cone of depression:

$$Q = P \cdot \frac{dy}{dx} \cdot 2\pi xy.$$

By integrating with respect to x and y

$$Q = \frac{\pi P y^2}{\log_e x} + C.$$

At the periphery of the island, the saturated thickness of water-bearing material is equal to H and at the wall of the well it is equal to h. The radii of the two cylinders so delineated are R and r, respectively. By substituting these values for x and y the constant of integration is eliminated and the following equation

is obtained

$$Q = \frac{\pi P(H^2 - h^2)}{\log_e R - \log_e r}.$$

This is the Dupuit formula for water-table conditions. The corresponding formula for artesian conditions is

$$Q = \frac{2\pi mP(H - h)}{\log_e R - \log_e r},$$

where m is the thickness of the artesian stratum.

These formulas presumably were developed for the purpose of predicting the yield of wells in areas where estimates could be made of the permeability of the water-bearing material. It appears unlikely, however, that the formulas were ever used generally for this purpose because the geologic and hydrologic conditions found in nature differ so widely from those on which the formulas are based. Wells are not situated in the centers of circular islands and therefore the cone of depression may expand almost indefinitely—that is, complete equlibrium probably will never be reached; some wells do not penetrate completely through the aquifer and some are not screened in all parts of their extent through the aquifer; the strata on which the aquifers rest are not generally horizontal; the aquifer generally differs from place to place in both thickness and in permeability; and the undisturbed water level is almost never horizontal. The Dupuit formula for water-table conditions is based on the erroneous assumption, moreover, that the water percolates to the pumped well in a horizontal direction, whereas, because of the drawdown of the water table, the movement is downward to some degree, and hence the cone of depression is altered somewhat from its theoretical form.[21] Despite the obvious difficulties in applying the formulas, they have been included for years in the hydrologic literature with little or no discussion of their limitations.

Thiem Formula.—In 1906 Günther Thiem [9] published the results of his investigation in connection with the development of an additional water supply for the city of Prague. In this investigation he determined the permeability of water-bearing ma-

terials by means of pumping tests, using the following modified form of the Dupuit formula:

$$P = \frac{Q(\log_e r_2 - \log_e r_1)}{2\pi m(s_1 - s_2)},$$

where P is the coefficient of permeability; Q is the rate of pumping, r_1 and r_2 are respective distances of two observation wells from the pumped well; m is the vertical thickness of the aquifer under artesian conditions, and the average vertical thickness of the saturated part of the aquifer at r_1 and r_2 under water-table conditions; and s_1 and s_2 are the drawdowns of the water level in two observation wells situated within the cone of depression. Thiem apparently was the first to employ the Dupuit formula for determining permeability and was also apparently the first to utilize the drawdown in two observation wells. In recognition of these innovations the formula is generally called the Thiem formula in the United States.

Limiting Formula.—In 1930 the junior author of this paper made a study of the theoretical aspects of the development of the Thiem formula in an effort to evaluate the effects of the discrepancies between assumed and field conditions upon computations of permeability. The study could not be carried very far, however, because very little was known of the behavior of the water table in the vicinity of discharging wells and a search of the hydrologic literature disclosed practically no experimental data on the subject. In order to obtain information of this kind a rather elaborate pumping test was made near Grand Island, Nebraska, in the summer of 1931. The drawdown of the water table during the period of pumping and the recovery of the water table after the pumping was stopped were observed through frequent measurements of the water levels in more than 80 observation wells situated out to about 1,200 feet on 8 lines radiating from the pumped well.[14]

The data obtained in the Grand Island test were given thorough study and a method of procedure was devised that takes into account the limitations in applying the Thiem formula to field conditions.[15] It was found that consistent results for permeability

can be obtained, provided the following procedure is observed: (1) Only those drawdowns are used that are obtained from observation wells situated on a straight line extending through the pumped well; (2) the drawdown of the water level s_1 is taken to be the average of the drawdowns on the opposite sides of the pumped well (preferably up-gradient and down-gradient) at the distance r_1 from the pumped well, and, similarly, s_2 is taken to be the average of the drawdowns on opposite sides of the pumped well at the distance r_2, etc.; (3) drawdowns are used only for wells within that part of the cone of depression that has reached approximate equilibrium in form by the end of the period of pumping; (4) drawdowns are used only for wells situated sufficiently far from the pumped well so that there are no appreciable effects on the slope of the water table from the failure of the well to penetrate the entire thickness of the aquifer, and from changes in permeability of the water-bearing material produced by the development of the well; and (5) the drawdowns obtained at more than two distances from the pumped well are used.

Successful determinations of permeability have since been made in Nebraska, Kansas, and elsewhere by following this empirical procedure. The formula now is sometimes called the limiting formula [25] and is written

$$P = 527.7qC,$$

where P is the coefficient of permeability as defined previously in this paper, q is the discharge of the pumped well in gallons a minute, and C is a constant, equal to A/B, that is determined graphically. A is equal to

$$\frac{\log_{10}\frac{r_2}{r_1}}{0.25M}$$

and B is equal to

$$\tfrac{1}{2}(s_{1u} + s_{1d} - s_{2u} - s_{2d}),$$

where r_1 and r_2 are distances, in feet, to two points on the cone of depression that lie on a straight line through the pumped well; s_{1u} is the drawdown, in feet, on the line at the distance r_1 up-

gradient from the pumped well; s_{1d} is the drawdown, in feet, on the line at distance r_1 down-gradient from the pumped well; s_{2u} is the drawdown, in feet, at the distance r_2 up-gradient; s_{2d} is the drawdown, in feet, at the distance r_2 down-gradient; and M is the sum of the saturated thicknesses of water-bearing material, in feet, at the locations of the 4 drawdowns.

To obtain C, all possible values of A are plotted against corresponding values of B and a straight line is drawn through the plotted points. More than one point is necessary, of course, to determine the plotted line and hence at least 6 observation wells —3 up-gradient and as many down-gradient—are required. C is determined from the slope of the straight line. If most of the points do not fall approximately on a straight line through the origin, the differences between assumed conditions and field conditions are thus indicated to be great, and the method cannot be successfully used.

Behavor of Ground Water in the Vicinity of Discharging Wells.—The pumping test made near Grand Island in 1931 and the several tests made since then in other localities have provided much factual information on the behavior of the ground water under different conditions in the vicinities of discharging wells.[18]

As soon as a pump begins discharging water from a well that penetrates an aquifer with a water table the water table is lowered around the well and a hydraulic gradient from all directions is established toward the well. The water table soon assumes a form which is comparable to an inverted cone, although it is not a true cone. Where the water-bearing material is homogeneous the base of this cone of depression, or area of influence, will be circular if the water table is initially horizontal, but somewhat elliptical if the water table has an initial slope. Some water-bearing material will be unwatered by the decline of the water table, and the water drained from this material will percolate toward the pumped well. Thus, for a short time, after pumping is begun, most of the water that is pumped may be obtained from unwatered sediments comparatively close to the pumped well, and temporarily very little water may be drawn to the well from greater distances. However, as pumping is continued, a hy-

draulic gradient that is nearly an equilibrium gradient will be established close to the well, and close to the well water will be transmitted through the water-bearing material at nearly the rate at which it is being pumped.

The decline of the water table and the resultant unwatering of material in this area will then proceed at a much slower rate, and water will percolate toward the well from greater distances. As the pumping of the well is continued, more of the formation will gradually be unwatered, an approximate equilibrium gradient that will transmit to the well approximately the amount of water that is being pumped will be established at increasing distances from the well, and an appreciable drawdown of the water table will be noted farther from the well. The water table near the pumped well, in order to maintain an approximate equilibrium form, will therefore, continue to lower indefinitely but at a decreasing rate.

If no water is added to the formation, the water table will continue to decline so long as the well is pumped and the cone of depression will eventually extend to the limits of the formation. Recharge to the formation, however, may halt the development of the cone of depression by furnishing additional water that will become a supply for the pumped well.

The piezometric surface under artesian conditions behaves in a manner very similar to the behavior of a water table. Water is at first removed from storage, not by the unwatering of a part of the formation but rather by the compaction of the aquifer and of the included and associated beds of silt and clay as the head is reduced. The squeezing out of water by compaction delays the development of the cone of depression in much the same way as the development of the cone of depression under water-table conditions is delayed by the unwatering of part of the formation. However, the quantity of water removed by compaction is generally much less than the quantity removed by the unwatering of a part of a formation, and consequently, the drawdown of the water levels in the observation wells and the development of the cone of depression is usually more rapid.

After the discharge of a well is stopped, water continues to percolate toward the well under the hydraulic gradient set up

during the period that the well was being operated, but instead of being discharged by the well it refills the well and the interstices of the material that were unwatered, or, under artesian conditions, it expands the aquifer and associated beds. As the formation near the well is gradually refilled, the hydraulic gradient toward the well is decreased and the recovery becomes progressively slower. At distances comparatively far from the well the water level may continue to decline for a considerable time after the discharge ceases, because at those distances water still is taken from the interstices of the material to refill the sediments around the well. In time there is a general equalization of water levels over the entire region and the water table or piezometric surface will assume a form similar to that it had under the original static conditions, although it may remain temporarily or permanently somewhat lower than before the water was withdrawn.

Gradient Formula.—The flow of water through the ideal hydraulic system postulated by Dupuit is equal to PiA, where P is the coefficient of permeability, A is the area of any one of a series of concentric cylindrical cross-sections of water-bearing material around the pumped well, and i is the hydraulic gradient at that section. Therefore,

$$P = \frac{Q}{iA} = \frac{Q}{2\pi ixy},$$

where y is the thickness of saturated water-bearing material at the distance x from the pumped well. The equation can be solved directly for P by determining the hydraulic gradient graphically from a profile of the cone of depression.

Where the water table, or piezometric surface, in its undisturbed condition has a slope, the direction of the flow of water toward a discharging well is not at right angles to the cylindrical sections around the well except along a line that extends directly up-gradient and down-gradient through the well. The flow through a given cross-sectional area at a specified distance up-gradient is greater than the flow through an equal cross-sectional area at the same distance down-gradient. The data collected in the pumping tests mentioned previously indicate that the hydraulic

gradient causing the flow toward the well is approximately equal to the average of the gradients at a given distance up-gradient and down-gradient from the well. Thus, for water-table conditions,

$$P = \frac{2Q}{\pi x(y_u + y_d)(i_u + i_d)},$$

where i_u is the hydraulic gradient and y_u is the saturated thickness of the water-bearing material at the distance x up-gradient, and i_d is the hydraulic gradient and y_d is the saturated thickness of water-bearing material at the distance x down-gradient.

The hydraulic gradient at any distance r on one side of the discharging well can be determined approximately from a profile of the cone of depression by (1) ascertaining from the profile the altitude of the water table or piezometric surface at the distances $(r + b)$ and $(r - b)$ from the pumped well and (2) dividing the difference in altitude of the water table at the two distances by $2b$. A study of the data collected in the pumping tests indicates that the hydraulic gradient can generally be determined satisfactorily from profiles of the cone of depression by taking the distance b as 10 feet.

For water-table conditions

$$P = \frac{18{,}335q}{r(y_u + y_d)(f_{(r+10)u} + f_{(r+10)d} - f_{(r-10)u} - f_{(r-10)d})},$$

and for artesian conditions

$$P = \frac{9{,}168q}{rm(f_{(r+10)u} + f_{(r+10)d} - f_{(r-10)u} - f_{(r-10)d})},$$

in which P is the coefficient of permeability as defined previously, q is the discharge of the well in gallons a minute, r is the distance, in feet, from the discharging well to any point on the part of the cone of depression that has reached an approximate equilibrium form, y_u and y_d are the respective thicknesses, in feet, of saturated water-bearing material at the distance r up-gradient and down-gradient from the pumped well, $f_{(r+10)u}$ and $f_{(r+10)d}$ are the respective altitudes, in feet, of the water level at the distance $r + 10$

up-gradient and down-gradient from the pumped well, $f_{(r-10)u}$ and $f_{(r-10)d}$ are the respective altitudes, in feet, of the water level at the distance $r - 10$ up-gradient and down-gradient from the pumped well, and m is the thickness of the aquifer, in feet.

A specific formula developed for the application of this method has been named the gradient formula.[25] It has been used successfully to determine permeability from the data obtained in pumping tests in Nebraska, Kansas and Arkansas.

Non-equilibrium Formula.—All the formulas so far described are based on the assumption that the hydraulic system can attain a state of equilibrium—a condition that is reached only approximately near the discharging wall. The factor of time is included in the formula only in the sense that the well is assumed to have been operating for a sufficient period to produce a state of equilibrium.

In 1935 Theis[16] published a formula for determining the drawdown of the piezometric surface in the vicinity of a discharging well, taking into account the removal of water from storage. This formula may be called the non-equilibrium formula because it does not depend on the hydraulic system reaching a state of equilibrium. The formula is based on the assumption that Darcy's law is analogous to the law of flow of heat by conduction and thus the mathematical theory of heat-conduction is largely applicable to hydraulic theory. The formula may be written

$$s = \frac{114.6q}{T} \int_{1.87r^2S/Tt}^{\infty} \frac{e^{-u}du}{u},$$

in which s is the drawdown, in feet, at any point in the vicinity of a well discharging at a uniform rate; q is the discharge of the well, in gallons a minute; T is the transmissibility of the aquifer (cubic feet of water discharged from each vertical column of aquifer with base 1 foot square for each foot of lowering in head); r is the distance, in feet, of the discharging well to the point of observation; S is a coefficient of storage, expressed as a decimal fraction; and t is the time, in days, that the well has been discharging.

The value of the integral

$$\int_{1.87r^2S/Tt}^{\infty} \frac{e^{-u}du}{u}$$

may be computed by the series

$$-0.577216 - \log_e u + u - \frac{u^2}{2.2!} + \frac{u^3}{3.3!} - \cdots,$$

in which

$$u = \frac{1.87r^2S}{Tt}.$$

The formula is based on the following assumptions: (1) the water-bearing formation is homogeneous, (2) the formation has an indefinite areal extent, (3) the discharging well penetrates the entire thickness of the aquifer, (4) the coefficient of transmissibility is constant at all places and at all times, (5) the discharging well has an infinitesimal diameter, (6) the undisturbed piezometric surface is horizontal, (7) the impervious bed underlying the aquifer is horizontal, and (8) water is taken from storage instantaneously with the decline in head.

Because the formula is based on the assumption that the transmissibility of the aquifer does not change during the period of discharge the formula can be applied approximately to artesian conditions. Where the thickness of saturated material is great, however, it appears likely that it can be used also for water-table conditions without introducing serious error. For water-table conditions, the coefficient of storage is the specific yield of the water-bearing material.

As T appears on both sides of the equation, the formula cannot be solved directly for transmissibility or the coefficient of permeability. However, T may be determined conveniently by the following graphical method suggested by Theis (personal communication, 1937).

The formula may be written

$$s = \frac{114.6q}{T} \cdot W(u),$$

in which $W(u)$ may be read "well function of u" and the other

terms are those previously defined. $W(u)$ is equal to

$$-0.577216 - \log_e u + u - \frac{u^2}{2.2!} + \frac{u^3}{3.3!} - \cdots$$

and u is equal to

$$\frac{1.87r^2S}{Tt}.$$

When T is to be determined from observations of the drawdown in one well, the logarithm of the drawdown is plotted against the logarithm of the reciprocal of time since the discharge began (s against $1/t$). When T is to be determined from the drawdowns in a line of wells, the logarithm of the drawdown is plotted against the logarithm of r^2/t. If the aquifer were entirely homogeneous and if water were discharged instantaneously with the decline in pressure, all plotted points (for all times and all wells) would fall on a smooth curve. The curve so determined is a segment of the type curve produced by plotting the logarithm of the value of the integral, $W(u)$, against the logarithm of the quantity u. If, therefore, (1) this latter curve (type curve) is plotted on logarithmic paper and (2) the observed drawdown values are plotted on transparent paper against $1/t$ (for one observation well) or r^2/t (for a line of observation wells) to a logarithmic scale the same as that used for plotting the type curve, then (3) the observed curve can be fitted to the type curve in only one place; (4) from this fit the value of $W(u)$ and the corresponding value of u may be determined from the type curve for any selected point on the curve of observed values, which (5) may be used in conjunction with the observed values for that point to determine T.

The transmissibility is then computed by the formula,

$$T = 114.6q\frac{W(u)}{s},$$

and the coefficient of permeability

$$P = \frac{114.6q}{m}\left(\frac{W(u)}{s}\right),$$

in which m is the saturated thickness of the aquifer.

This method for determining permeability has been successfully applied to pumping tests made in Nebraska and elsewhere.[25] The coefficients of permeability determined from the drawdowns in wells located on a line through the pumped well (averaged in the same manner as for the limiting formula) checked very closely the permeabilities computed by both the limiting and gradient formulas.

Recovery Formula.—Theis introduced also a formula for determining transmissibility from the recovery of the water level in a well after its discharge has stopped.[16] This formula is based on the assumption that if a well is allowed to discharge for a known period, the residual drawdown at any instant after the discharge of the well has stopped will be the same as if the discharge of the well had continued but a recharge well with the same flow had been introduced at the same point at the instant the discharge actually stopped. The hydrologic and geologic assumptions are the same as for the non-equilibrium formula.

The formula may be written

$$T = \frac{264q}{s} \log \frac{t^1}{t},$$

where T is the transmissibility; q is the discharge of the well in gallons a minute; s is the residual drawdown of the water level, in feet; t^1 is the time since the discharge started, expressed in any unit of time; and t is the time since the discharge stopped, in the same unit.

The value of log $(t^1/t)/s$ should be determined graphically by plotting log t^1/t against s. If most of the points do not fall on a straight line the formula cannot be applied with assurance.

Extent of Cone of Depression.—Until very recently little attention was given to the effects of the length of the period of discharge on hydraulic problems relating to wells. The time that wells are pumped or allowed to flow necessarily enters almost all computations of ground-water flow, but in many cases it has been included implicitly—usually in an assumption. In formulas of the equilibrium type only the ultimate condition of the ground-water system is considered; thus the period of discharge is as-

sumed to be infinite, and no consideration is given to the differing hydraulic conditions that exist prior to stability.

The disregard for the non-stable conditions of ground-water flow probably has been due, at least in part, to the difficulties involved in mathematical treatment. The recent introduction of the non-equilibrium method by Theis [23] has materially aided the formulation of correct concepts, but precise mathematical treatment of all factors involved in a non-stable system has not as yet been accomplished.

Observations on the behavior of the water table around pumped wells made in connection with the pumping tests conducted by the Geological Survey and cooperating parties show that the basic assumption of the equilibrium-type formulas, namely, that equilibrium has been reached, is for practical purposes valid for only a small area around a pumped well. Beyond this small area the assumption is far from true.

There is, of course, an appreciable drawdown of the water level far beyond the area in which the cone of depression has attained a form that is essentially an equilibrium form. The pumping tests indicate that the cone of depression reaches essential stability in form to a distance from the pumped well that is only a small proportion of the distance to which the effect of the pumping is transmitted. This observed condition is significant because it definitely limits the rigorous use of the equilibrium-type formulas to a very small area around a discharging well and virtually invalidates the use of the formulas to larger areas unless the period of discharge is very long.

Many of the formulas of the equilibrium type given in the literature, including those of Dupuit,[4] Slichter,[7] and Turneaure and Russell,[8] include the determination of R, the distance from the discharging well at which the drawdown of the water level is inappreciable. Such formulas also involve the assumption that a condition of equilibrium exists over the entire area of influence —that is, from the discharging well to the distance of R. As has just been pointed out, this assumption is far from correct. Several investigators have given arbitrary values to be used for R: Slichter,[7] 600 feet; Muskat,[21] 500 feet; and Tolman,[20] 1,000 feet.

Although it is obvious that the use of R in the equilibrium-type formula will generally result in determinations that are more or less in error, criticism of its use is based probably more on the implication that R actually represents the distance from a discharging well at which the effect of the discharge is negligible. Leggette [22] recently observed appreciable fluctuations of water level in wells caused by the shutting down of pumped wells situated 1 to 7.1 miles distant. The extent of the cone of depression is of very practical significance in determining the spacing of wells and in the solving of many quantitative problems. It also has been the crux of important legal controversies. Because small empirical values for R, presumably intended chiefly for the solving of formulas for the discharge of wells in areas of known permeability, appear so persistently in the literature, it has generally been assumed that the extent of the cone of depression is actually small. That the cone of depression may extend far beyond 500, 600 or 1,000 feet is, however, a correct theoretical deduction and has been demonstrated by field observations.

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THE SPODUMENE PEGMATITES OF NORTH CAROLINA.[1]

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THE pegmatites of North Carolina are legion and elsewhere in the Atlantic states, from South Carolina to Maine, are many more, but outside of a few scattered masses in New England and the remarkable belt in North Carolina to be described in this paper, no spodumene-bearing pegmatites are known in the great expanse of granite exposed in these states.

As is well known, spodumene is found only in the granitic pegmatites and in comparatively few of those. Except for an occurrence in Trans-Baikalia, of which little has been made known, and another in South Africa,[2] spodumene is known in large quantity only in the Western Hemisphere. In South America large deposits have been reported from Argentina[3] only, and so far no description of them is known. In North America there are deposits in Colorado, Manitoba, New England, New Mexico, North Carolina, and South Dakota, but the greatest of these is in North Carolina.

Other lithium minerals are known in the great granite areas of Africa, Asia, and Australia, and it would not be surprising if other deposits of spodumene were found in these continents.

The pegmatites of North Carolina have long been known for their rare minerals, for their feldspar and mica, and for their numbers. As pegmatites are formed from the solutions squeezed from freezing magmas it is probable that a similar suite of unusual minerals carried by adjacent units or groups indicates that

[1] Published by permission of the Director, Bureau of Mines, U. S. Department of the Interior. (Not subject to copyright.)

[2] Gevers, T. W., Partridge, F. C., and Joubert, G. K.: The pegmatite area south of the Orange River in Namaqualand. Union S. Africa Geol. Surv. Mem. 31: 157–160, 1937.

[3] In the departments of San Martin and Chacobuco. Gerez, José M.: Estadistica minera de la nación, Año 1936. Dir. Min. & Geol. Arg. 1937, pp. 17–18.

they have come from the same magma and those carrying a different suite have come from another magma.

Judged by this standard alone, the pegmatites of North Carolina fall into the following groups: Pegmatites that contain (1) zircon and sphene, (2) uranium and rare earths, (3) lithium and tin, and (4) simpler granite pegmatites that carry some beryl. The last group is really several rather than one, and simple unmetamorphosed pegmatites may occur with any group.

GENERAL GEOLOGY.

In general the rocks of the North Carolina spodumene pegmatite belt are biotite-quartz schists or muscovite-quartz schists lying in broad folds approximately parallel to the trend of the Appalachians. They have been intruded by various granites that have brought with them numerous pegmatites into one group of which lithium (mostly as spodumene) and tin were later introduced. The whole area has been eroded into the fairly level Piedmont, a peneplain above which stand great masses such as The Pinnacles, Henry Knob, Nannies Mountain, and others that have been described as quartzites but which are apparently schists silicified and otherwise altered by magmatic solutions.

In a country with such a geologic history there have been, of course, many movements that have cracked, crushed, folded, and faulted the rocks. The country has been exposed to weathering for an immense length of time and the rocks are greatly but unequally decayed. In some places on the Piedmont decay has gone downward more than 100 feet. In other places the rocks on the surface are remarkably fresh.

In general, the spodumene pegmatites, which generally are also the tin-bearing pegmatites, are resistant to weathering and their outcrops stand well above the surface, but in places the pegmatites have decayed like other rocks and the partly decomposed spodumene can be found in the residual clay. At the Lincolnton tin mine, veins of halloysite as much as 2½ inches thick have formed in the weathered pegmatite.

The parent granite of the lithium-tin pegmatites is the White-

side granite of Keith and Sterrett,[4] and is possibly of late Carboniferous age. It is a fine-grained, somewhat gneissoid rock containing both muscovite and biotite. The biotite is variable in quantity and in places is not visible to the unaided eye.

The granite is greatly decayed where exposed 2½ miles southwest of Kings Mountain, but is comparatively fresh on Long Creek and in its vicinity, 3 to 4 miles north of the town. It occurs in considerable quantity 1 to 2 miles south of Long Shoals, where it is far from fresh, carries considerably more biotite, and is noticeably gneissoid. Other outcrops I have not seen.

Many pegmatites follow the course of the granite but only a part of them carry spodumene.

THE SPODUMENE PEGMATITES.

Distribution.—Cassiterite-bearing pegmatites have been known in North Carolina for many years, have been repeatedly described and are now being studied again by the United States Geological Survey. Although J. Lawrence Smith[5] described emerald-colored spodumene from eastern Alexander County in 1881, the first mention that I have seen of the vastly more plentiful but quietly colored spodumene in the North Carolina pegmatites was that of Graton[6] in 1906, 25 years later. After another 25 years the spodumene was mentioned by Keith and Sterrett[7] but neither publication emphasized the occurrence and the deposits remained almost unknown. The flood of geological publications is too great for anyone to digest, and unless such items are emphasized, they are apt to be unnoticed.

The spodumene-bearing pegmatites, and probably a larger number of the same age and parentage that do not carry spodumene, lie in a belt extending from Grover on the North Carolina-South

[4] Keith, Arthur, and Sterrett, Douglas B.: Gaffney-Kings Mountain Folio, South Carolina-North Carolina. U. S. Geol. Surv. Geol. Atlas of the U. S. No. *222*: 6, 1931.

[5] Smith, J. Lawrence: Hiddenite, an emerald green variety of spodumene. Am. Jour. Sci., *21*, ser. 3: 128–130, 1881.

[6] Graton, L. C.: Reconnaissance of some gold and tin deposits of the southern Appalachians. U. S. Geol. Surv. Bull. *293*: 128–130, 1906.

[7] *Op. cit.*, p. 11.

Carolina line to Lincolnton about 25 miles northeast (Fig. 1). From Grover to Kings Mountain, the main line of the Southern R. R. follows closely on the west side of the belt; north of Kings Mountain it is on the east side of the belt, but at no place is either the main line or a branch far distant. The deposits are thus

FIG. 1. Map showing location of belt of spodumene pegmatites in North Carolina.

within easy reach of the principal industrial centers of the United States.

On the south, the belt ends at the state line in a single pegmatite which, as exposed, seems to be about 90 feet wide but may be wider. The next known deposit is about a mile to the northeast, on the Westmoreland farm, where a pegmatite between a quarter

and a half-mile long and 6 to 60 feet wide appears, and within 200 yards on the east are a number of satellites.

Between here and the W. A. Ware farm 6 miles farther to the northeast, there is a space of about 2 miles in which are numerous barren pegmatites, but none are known that contain spodumene. For the remaining 4 miles there are spodumene-pegmatites most of the distance. Several are from 600 to more than 1000 feet long with widths ranging from 10 to more than 100 feet.

At one place there are five parallel pegmatites within 250 feet and at other places there are two parallel pegmatites. On the Ware farm 2½ miles southwest of Kings Mountain and for a mile to the north, the spodumene pegmatites have reached their greatest development; several pegmatites spread out in long radial arms and there are several rounded masses. L. M. Williams has been working in a pegmatite that is 76 feet wide and is apparently more than 1,000 feet long. Several inclusions of schist have been exposed, and there are probably such inclusions in all of the wide pegmatites.

From the south side of Kings Mountain to Long Creek Church, about 4 miles north, only comparatively small spots of the spodumene pegmatite are known. In the town a pegmatite 8 feet wide is exposed near the north end of the west side of the railroad cut, and others are exposed a short distance east. Within the town limits tin has been mined, although unprofitably, from two pegmatites of the belt but no spodumene is known to have been found in them.

On Long Creek near the church of that name is a good exposure of the parent granite, near which are several moderately sized spodumene pegmatites and on the Sarvice farm half a mile west and northwest of the church the granite outcrops broadly and is accompanied by a group of pegmatites in the schist, one of which is about a quarter of a mile in length and more than 100 feet in width. On it, an experimental kiln for the decrepitation of spodumene has been built. From the Sarvice farm to the last known outcrop beyond the Lincolnton tin mine (which is also in the spodumene pegmatite belt) there are many more spodumene pegmatites scattered through a width of three miles.

A tin-bearing pegmatite on Beaver Dam Creek about 6 miles southwest of the tin mine seems to be considerably outside of the main part of the belt, but carries some spodumene. Both pegmatite and enclosed spodumene are greatly decayed. Another spodumene pegmatite is reported as near Lithia Inn, a mile or more north of the old tin mine.

The hiddenite pegmatite in Alexander County is in line with the belt but the occurrence is peculiar and will be referred to again.

Most of the pegmatites follow the strike of the schists but the general direction of the group seems to be somewhat nearer north than the structure of the country, and, although exposures are poor, in some places where the pegmatites strike in various directions, it seems probable that they cut across the schistosity.

As has been said, it appears that the spodumene pegmatites are part of a much larger group of pegmatites. In the beginning they were masses of moderately fine-grained granitic minerals (that is, they were fine grained when considered as pegmatitic minerals) —quartz, potash feldspar, and so much muscovite that many of the less metamorphosed units constantly tempt exploitation, although the mica plates are small, ordinarily ½ to 2 inches broad, and would be fit only for grinding. If, as appears probable, the pegmatites originated from the biotite-bearing Whiteside granite, it seems remarkable that no biotite is found in them. The spodumene and other lithium minerals were introduced along cracks that affected only part of the pegmatites.

The cracking that preceded the flow of the lithium-bearing solutions seems to have afforded channels that carried part of the solutions far afield and this long-distance movement seems to account for the hiddenite found in Alexander County. The pegmatite that contains the hiddenite is of a different type from those between Lincolnton and Grover. It is in narrow bodies from a fraction of an inch to a few inches wide and where now visible is made up mostly of white microcline that has been in large crystals.

The microcline is now netted with narrow crushed zones, between which are cracked kernels, part of which are almost glassy clear (Fig. 2). Qualitative spectrochemical tests on the clearest

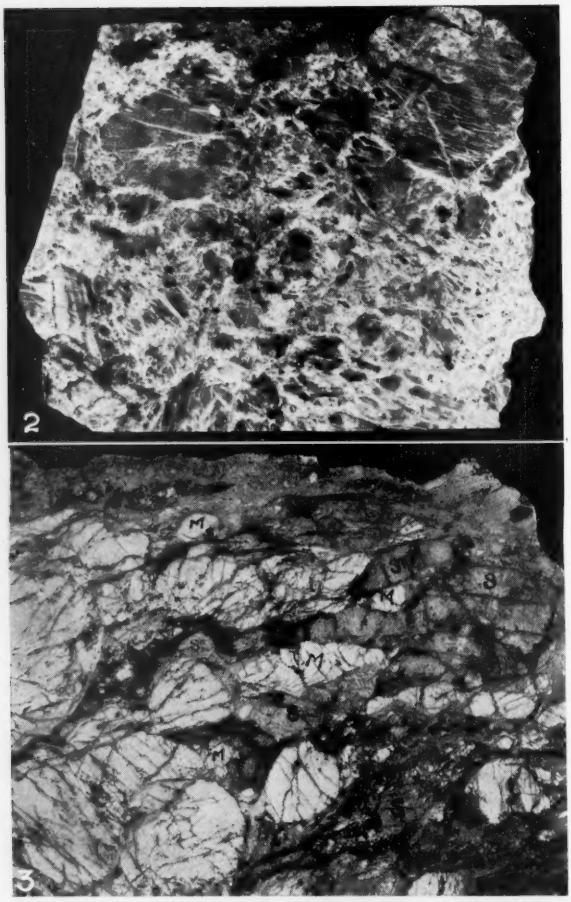


FIG. 2. Thin section of pegmatite from Hiddenite, North Carolina. Crushed clear microcline with tourmaline and biotite (rough edges) introduced in crushed lines. Abt. 3¼ × 4 inches.

FIG. 3. Pegmatite with fragments of microcline (M) rounded by solution then replaced partly or wholly by spodumene (S). From the Ware farm 2½ miles SW of Kings Mountain, N. C. Polished slab about 8¾ × 11 inches.

microcline showed about 7 per cent of both K_2O and Na_2O, and a quantitative spectrochemical determination gave 0.88 per cent Rb_2O. No Li or Cs could be detected.[8] Plagioclase is present in microscopic tear-shaped inclusions.

In the crushed lines are small grains of quartz, tiny flakes of muscovite, black tourmaline, and a little leucoxene and biotite. Palache, Davidson, and Goranson[9] mention numerous other minerals, but in the dump material that I examined they were either not present or in minute quantity only. The hiddenite is largely in small vugs and, like the spodumene in the other pegmatites, is later than the microcline. The green color is probably to be explained by the passage of the lithium-bearing solution through cracks in dunite or other chromium-bearing rocks from which the green-forming element was dissolved to be combined with the precipitating spodumene.

Tiny pegmatites, crumpled like an entrail (Fig. 4) are supposed by the authors[10] quoted to have been given their visceral form by the folding of the enclosing schist, but the schist shows no similar folding and parallel dikelets show no parallelism of folding. Squeezing of the molten material into potential cracks in a gradually and unevenly yielding schistose rock seems to offer a more plausible explanation.

In the same way that the lithium-bearing solutions were carried far to the north in channels formed by movements and cracking that preceded lithium deposition, so tin-bearing solutions were carried to the south through cracks formed by movements preceding the deposition of cassiterite, and the tin-bearing pegmatite was formed at Gaffney, South Carolina, 10 miles southwest of the terminal (so far as we now know it) lithium deposit at Grover.

On the other hand, between the Westmoreland place and the Detmer place, a distance of 2 miles, both tin and lithium solutions lacked channels to reach the pegmatites, but a later flow of barren

[8] All spectrochemical determinations given in this paper are by Morris Slavin of the Eastern Experiment Station of the Bureau of Mines, for whose help I here note my grateful appreciation. All figures quoted should be credited to him.

[9] Palache, Charles, Davidson, S. C., and Goranson, E. A.: The hiddenite deposits in Alexander County, North Carolina. Am. Min., 15: 280–302, 1930.

[10] *Op. cit.*, p. 283.

solutions found openings to carry large quantities of quartz into this stretch.

The extra-terminal deposits of tin and lithium and the barren stretch in the belt of lithium-bearing pegmatites indicate that the pegmatite-forming solutions had very different characteristics at different times, and they also give some indication of the fact that a movement in the rocks during the flow of the pegmatite-forming solutions is often, if not always, accompanied by a change in the

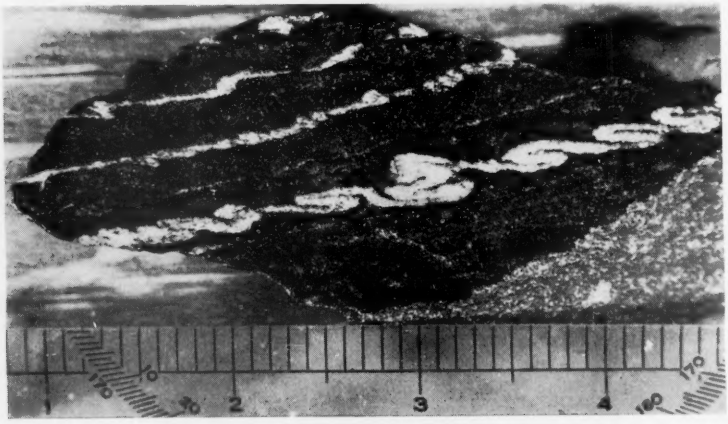


FIG. 4. Visceral pegmatite dikelets. Hiddenite, North Carolina. 1/32–1/8 inch thick.

constitution of the solutions.[11] The events are of course concomitant. Gradual change in temperature and consequent change in volume, and possible renewal of intrusion are reflected in change of content of solutions, replacement, and deposition.

Effects.—Here and there the biotite of the schists has been developed next to the pegmatites into larger plates and has been

[11] A concept denied by some European geologists; thus A. Fersman (On the regularity (law) of the composition of granitic pegmatites. Compt. Rend. Acad. Sci. U.S.S.R., No. 4, 1931, pp. 115–122) says "the pegmatitic complex must be considered a closed system, and therefore, its composition, course of the process, etc., result from the physico-chemical laws but not from the assumption of changes in the composition of the passing solutions, intrusion of new emanations, etc."

segregated into small rich masses by the metamorphic action of the pegmatitic solutions. This is a more or less common phenomenon,[12] and here the rare alkalis have been introduced into the mica.[13] M. C. Hardin sent to me specimens of weathered biotite schist from a body lying between two spodumene pegmatites on the Westmoreland place, near Grover. The biotite plates show no noticeable development but on a general sample of the schist spectrochemical analysis gave the following components and percentages: Li_2O 3.2, Rb_2O 0.44,[14] Cs_2O 0.19, Na_2O 0.96, K_2O 5–10, MgO 5–10 — a rock that in case of necessity could readily be used as a source of lithium.

In places, the muscovite in the schist also shows some development next to the pegmatite, but it seems to have acquired none of the rare alkalis.

GEOLOGIC HISTORY.

The pegmatites are of the metamorphic type; that is, the original minerals of the pegmatite have been altered or replaced by later solutions, as is true of all pegmatites carrying lithium minerals and probably all other unusual minerals. Along broad lines, at least, the history of the group seems to be rather plainly recorded. The pegmatites appear to have grown by the following steps:

(1) The original pegmatites were formed by solutions forced from the freezing Whiteside granite after its intrusion along the strike of the Piedmont schists, and in general the pegmatites were formed in cracks that also followed the strike of the schistosity.

(2) After the pegmatites were formed movement took place along the original cracks and the pegmatites were sheeted parallel to their lengths.

(3) At many places a white microcline, in part not visibly perthitic, then replaced much of the original pegmatite. The proportionate introduction of the microcline into the various pegmatites was very unequal

[12] See Vogt, J. H. L.: The physical chemistry of the magmatic differentiation of igneous rocks. III, Second half, Skrif. Vid.-Akad., Oslo, 1930. I. Mat.-Nat. Klasse, No. 3, 1931, p. 84. Vogt also quotes Tom Barth on similar observations.

[13] Hess Frank L. and Stevens, Rollin E.: A rare alkali biotite from Kings Mountain, North Carolina, Amer. Min., *22*: 1040–1044, 1937.

[14] The figures for Rb and Cs are reliable only when compared between themselves, as at this time no accurate standard has been obtainable. The figures are, however, probably as nearly correct as most of the published determinations of the two elements.

(4) Later a second movement took place along the old cracks and the microcline was crushed and broken.

(5) Following the second movement, solutions carrying sodium, lithium, and probably aluminum, as well as the ubiquitous silicon, rose along the cracks and filled the spaces between the fragments of microcline with a fine-grained mass of albite, quartz, spodumene and a little muscovite, making a secondary albite aplite. The aplite not only filled spaces but replaced the pegmatite in places, probably where the pegmatite had been closely sheeted and finely broken, replacing considerable lengths and widths so that the masses now look as if they had a flow structure, and fragments of microcline were rounded like waterworn pebbles. These were replaced, some wholly, some in part, by spodumene (Fig. 3) and crystals of spodumene replaced both early pegmatite, if any is left, and the later microcline. In places crystals 8 or 10 inches long extend outward in each direction from cracks that formed solution channels.

(6) Immediately after the spodumene had been deposited parts of the pegmatites were greisenized by solutions that carried tin, and cassiterite was introduced into or near the greisen.

(7) In places a third cracking occurred.

(8) The third period of cracking was followed by a flow of solutions carrying large quantities of silica that formed masses of quartz most of which are less than a foot thick, and in many of which the quartz is in small, distinct, clear grains. This may be seen particularly on the Detmer place, 4½ miles southwest of the Kings Mountain town line, and in a group of pegmatites about a mile southwest that show no spodumene. Some of the quartz, as on the Westmoreland farm, is very glassy. The quartz replaced the spodumene extensively, and in places seems to have been followed by amblygonite, found in small quantity on the W. A. Ware place.

(9) The pegmatite was fractured a fourth time and cracks in the latest quartz and the amblygonite were filled by zincblende and quartz (Fig. 3). The zincblende is a dark brownish red and carries between 1½ and 2 per cent of iron as determined spectrochemically. It probably came in after the solutions had cooled greatly.

Unlike many metamorphic pegmatites, few large masses of nearly pure feldspar, quartz or spongy mica have been formed, but in the old tin workings 2½ miles southeast of Lincolnton were modest bodies of feldspar (mostly kaolinized) and of spongy mica. Other bodies of kaolinized feldspar seem to have been present in the two old tin mines in the north and south parts, respectively, of the town of Kings Mountain. If, during the development of the pegmatites, steps four and five [(4) the cracking of the microcline bodies and (5) their replacement by spodumene and accompanying minerals] had not taken place, there would have been, of course, considerable masses of rather pure feldspar and it seems possible that there may have been such

a hiatus in the flow of solutions through the pegmatite once mined for tin in the north part of Kings Mountain.

PEGMATITE VEINS.

At Long Creek church, on the Sarvice farm ½ mile northwest of the church, and on a hill 1½ miles south of Long Shoals, pegmatites of spodumene, quartz, and microcline with subsidiary green apatite and possibly amblygonite, cut the granite. They follow the observations of Vogt[15] and Andersen[16] that most pegmatites in the parental granite are much smaller than those in the adjacent rocks.

These pegmatites are narrow, partly linear bodies, but at Long Creek they have sudden irregularities of shape and breadth. The bodies seen are 3 to 12 inches wide where straight, but have wider irregularities and are not more than 25 feet long. The first two mentioned are vertical but the third is apparently flat. In mineral composition they are much like the spodumene pegmatites of the region but the disposition of the minerals is so strikingly different that it seems worth while to give them more attention than their economic value warrants.

Through the generous cooperation of L. M. Williams of Charlotte, N. C., large specimens were quarried from the Long Creek occurrence and shipped to Washington for study. The vein is 8 inches wide. One specimen showed a length of about 20 inches horizontally along the vein and another more than twice as much.[17] Both specimens have from 4 to 6 inches of granite on each side of the vein. Sections have been sawed from the end of the smaller piece and the remainder has been sliced horizontally (Figs. 5 and 6).

The granite shows a somewhat gneissoid structure that forms an angle of about 45° with the vein on one side and of about 30° on the other side, when seen in vertical section (Fig. 6). The

[15] Vogt, J. H. L.: The physical chemistry of the magmatic differentiation of igneous rocks. III, 2nd half. Norske Vidensk.-Akad., Oslo, Skr. I Mat.-Nat. Kl., No. 3: 56, 1930.

[16] Andersen, Olaf.: Discussions of certain phases of the genesis of pegmatites. Norske Geologisk tids., Bd. 12: 6–7, 1931.

[17] Now in the United States National Museum.

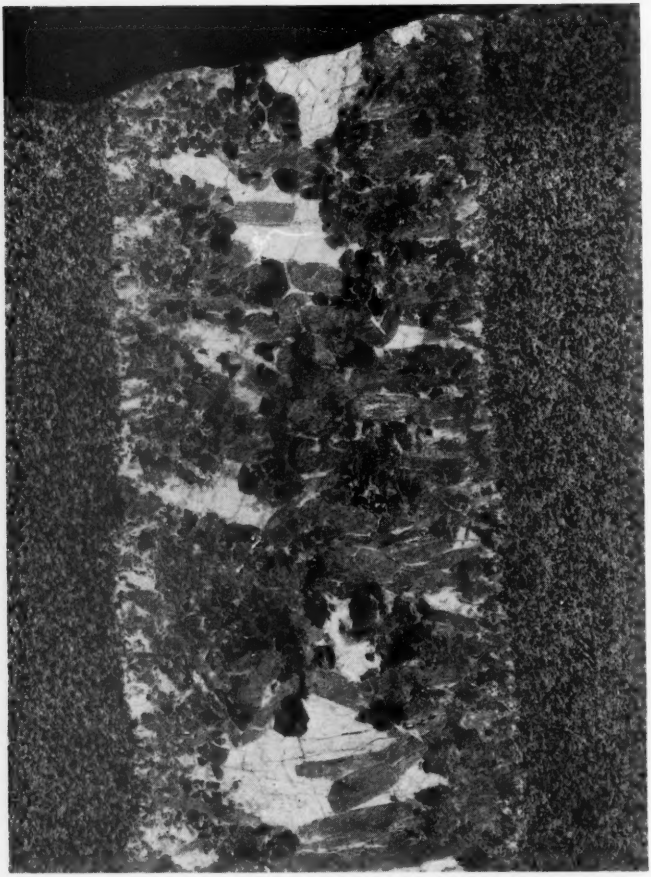


FIG. 5. A pegmatite vein cutting granite. The nearly white masses are microcline, gray crystals are spodumene, quartz is darker, and muscovite nearly black. Polished horizontal slab, vein 8 inches wide. From Long Creek Church, 3 miles north of Kings Mountain, North Carolina.

gneissoid structure is not visible in a horizontal section (Fig. 5). For an inch or two from the vein the muscovite is noticeably darkened. The muscovite of the granite is ordinarily slightly pleochroic, probably due to a small content of iron and the darkening near the pegmatite seems to mean a greater content. Unfortunately, the granite like other rocks of the region has been

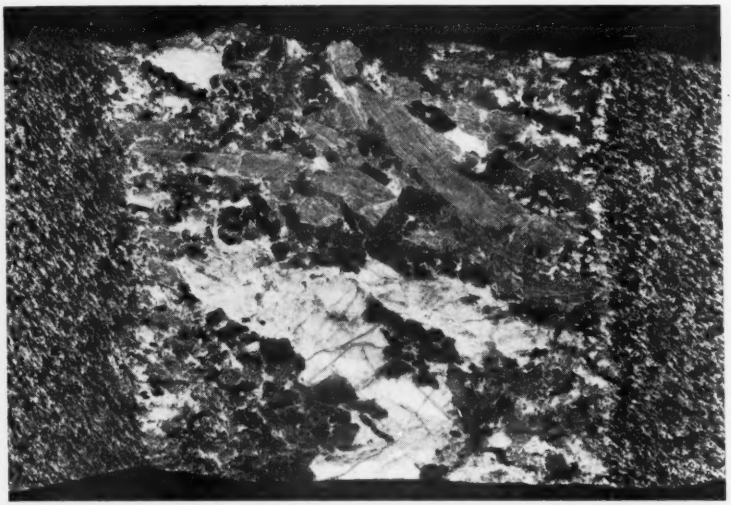


FIG. 6. Vertical section of pegmatite shown in Fig. 5. Note the spodumene crystal starting at right center and changing its direction of growth, conforming to the gneissoid structure of the two sides. A crystal on the left side shows a like tendency.

exposed to weathering for an immensely long time and nearly all of the muscovite is more or less stained and coated with iron oxide, so that it cannot be certainly determined whether the micas contain more or less than their normal quantity of iron, because, if treated with acid to remove the extraneous iron, that belonging to the mica may also be attacked. There is no apparent silicification of the granite walls.

The vein does not have smooth walls but crystals of slightly

greenish yellow spodumene and grayish quartz that bristle inward are bluntly rooted in the sides. On the walls the crystals of spodumene are mostly an inch or less long but some are 6 inches long and an inch thick.

In Fig. 6 a long crystal of spodumene is seen to start from the right side in about the direction of the gneissoid structure on that side. The crystal bends upward and takes the direction of the more steeply inclined structure on the left side, making it seem probable that movement was going on as the crystal formed. Another crystal growing from the left side bends similarly, but in the opposite direction. In horizontal sections such crystals are cut through and appear to be normal to the sides but to reach neither side. There are many more crystals on one side than on the other. Most of the apatite crystals are nearer the same side and small cavities from which amblygonite appears to have weathered, range 1 or 2 inches from the same side. Here and there are masses of somewhat perthitic white microcline. In the vertical cross-section of the vein (Fig. 6) the masses have an elongation roughly alined with the gneissoid structure of the granite. In the horizontal section no regularity can be seen but the feldspar is pierced by spodumene and quartz crystals and has quite evidently been replaced by them. The replacement is much more evident in the vertical cross-section.

The history of these pegmatites seems to have been simpler than that of the pegmatites in the schist: (1) Cracks were formed in the granite, probably while the movement that made it gneissoid was taking place, and (2) the cracks were filled with white microcline (stage 3, of origin), evidently from solution, because the granite was replaced at the same time, as shown by sudden irregularities in the pegmatite and the rough frozen sides. These, apparently, are pegmatites to which the word "veins" can be properly applied. (3) Another movement, in which the sides moved unevenly as the granite was given further gneissoid structure, cracked the microcline diagonally and was accompanied by solutions that replaced the feldspar by spodumene, quartz, and a little muscovite, with less amblygonite and still less green apatite. (4) Later cracks more nearly normal to the sides allowed quartz

to replace microcline, spodumene, and mica and some of the apatite was carried into the granite 2 inches or more from the pegmatite. It is notable that some of the cracks were only partly healed. I found similar partly healed cracks along which siliceous solutions had replaced spodumene and other earlier minerals at Lithia, Mass.

Spectrochemical estimates of the alkalis in the microcline and mica of the veins gave the following results:

	Li_2O	Rb_2O	Cs_2O	Na_2O
Muscovite	None	0.27	None	1.4
Microcline	2.6	0.23	None	3.4

RARE ALKALIS IN THE GRANITE.

Owing to the configuration of the ground it was impracticable to get a fairly fresh piece of granite more than about 18 inches from the pegmatite, though it would have been desirable to have gotten it as many feet away.

The granite is fine-grained, the grains possibly averaging 1/16 inch in diameter. At this place the biotite forms only about 10 per cent of the mica or 0.5 per cent of the granite, of which the feldspar forms possibly 60 per cent. James Norman kindly separated the feldspar and mica from the quartz and from each other by flotation, and I separated the micas from each other electromagnetically. Perhaps 5 per cent of quartz remained in the feldspar fraction and 3 per cent of feldspar and quartz in the mica. After magnetic separation the micas seemed to be very clean though there was undoubtedly some contamination. Spectrochemical estimations of the alkalis in the fractions gave the following figures:

	Na_2O	Li_2O	Rb_2O [18]	Cs_2O
Orthoclase	heavy	n.d.	0.8	n.d.
Muscovite	0.6	0.9	1.2	n.d.
Biotite	0.8	1.0	1.4	0.5

[18] See footnote 14.

Where " n.d." is shown, the quantities were not more than 0.1 per cent. According to these determinations here is a granite that contains in the neighborhood of 1 per cent of rubidia. Of the granitic minerals, the biotite carries the largest percentage of the rare alkalis. Farther north the granite carries much more biotite and if the biotite contains a similar percentage of Rb_2O the granite should be considerably richer. The quantity of rubidium in the area seems to be very large. Some day it may be wanted—stranger things have happened.

DETAILS OF THE MAIN PEGMATITES.

The main spodumene pegmatites are of comparatively even texture and as stated, carry very little muscovite, and that only in small books. There are remarkably few minerals other than feldspar, quartz, muscovite, and spodumene. A little black tourmaline that is blue in thin section, a much smaller quantity of green tourmaline, very few garnets, scattered dots of green apatite, and scarcer specks of columbite are found. A few small white beryls including one of the colorless goshenites have been seen. In 1904 Graton [19] obtained at the Faire's tin mine at the south edge of the town a new manganese phosphate that he called purpurite, but not another specimen was found in the entire belt until 1938. It is probably a weathering product from lithiophilite a little of which was also found with the original purpurite, but none has been found since. A single piece of cookeite was found by L. M. Williams 3½ miles southwest of Kings Mountain.

In examining spodumene concentrated by flotation, Dr. Alton Gabriel found particles of amblygonite that apparently had been enclosed in the spodumene crystals. A little later, L. M. Williams found pieces of 2 or 3 cubic inches on the W. A. Ware place, and after that pieces were found at other points (Fig. 7). As already said, cavities in the spodumene of the pegmatite veins on Long Creek evidently have been formed by the removal of amblygonite through weathering.

[19] Graton, L. C. and Schaller, W. T.: Purpurite, a new mineral. Am. Jour. Sci., 20, ser. 4: 146–151, 1905.

Cassiterite has been found at numerous places and has caused the expenditure of a considerable amount of money in profitless efforts at mining. As already noted, most of the cassiterite occurs where the pegmatite has been greisenized, and the mica of

FIG. 7. Zincblende replacing quartz and amblygonite (A). Size of specimen 3⅝ × 5⅜ inches (9.2 × 13.7 cm.). From pegmatite 2½ miles southwest of Kings Mountain, North Carolina.

the greisen would itself indicate that spodumene was not present, at least in that part of the pegmatite. In most places where tin has been found without greisenization, the pegmatite is greatly decayed. Here and there specks of cassiterite are found in spodumene, and it seems probable that tin was brought in either by a late phase of the lithium-bearing solutions or by a later separate

solution. This is the case in the Etta pegmatite in the Black Hills of South Dakota where cassiterite is found on spodumene crystals or in cracks in them.

Not the least remarkable mineral is the feldspar. As already noted, some shows almost no perthitic structure but part shows an ordinary pattern. Spectrochemical tests on specimens of the feldspar, spodumene, and muscovite for the alkalis gave the following results:

	K_2O	Na_2O	Li_2O	Rb_2O [20]	Cs_2O
Microcline 1½ mi. NE. of Grover, N.C.	10+	5–10	0.2	1–2	0.1
Microcline 1½ mi. NE. of Grover, N.C.	10+	5–10	trace	1–2	0.1
Microcline W. A. Ware farm, Kings Mt., N.C.	10–15	2–3	trace	1–2	0.2
Spodumene W. A. Ware farm, Kings Mt., N.C.	0.5	0.5	5–10	trace	none
Muscovite, mixed, from several pegmatites	n.d.	1.5	0.37	1.3	none

[20] See footnote 14.

The microcline from the W. A. Ware farm shows a soda content of the order of magnitude common to typical perthitic microclines.

Norman and Gieseke [21] in describing flotation experiments on the spodumene say: "The tailings contained much less spodumene than their Li_2O content indicates, because of the lithium-bearing mica, which is not floated with the spodumene. The tailings from . . . high-grade rock contained 0.55 to 0.65 per cent Li_2O when only about 1 per cent spodumene was found to be present by microscopic examination." These figures seem to indicate a somewhat higher Li_2O content of the mica than was shown spectrochemically.

The spodumene crystals vary considerably in size from microscopic size upward (Fig. 8), although there are no big crystals like those of the Black Hills. The largest crystals seen (from the W. A. Ware place) are about 2½ feet long (Fig. 9). Probably the average sized crystal is no larger than one's finger, but like diatoms they make up in numbers what they lack in size, and large parts of the pegmatites carry 15 per cent or more spodumene.

[21] Norman, James, and Gieseke, E. W.: Beneficiation of spodumene rock by froth flotation. A. I. M. E. Tech. Pap. 1161: 2, 1940.

Nowhere else is there known to be a group of pegmatites in which the spodumene is spread so consistently through the whole mass as in the Kings Mountain belt. Out of hundreds of chips broken at random from the lichen-covered rocks, I doubt if 2 per cent have not shown visible spodumene.

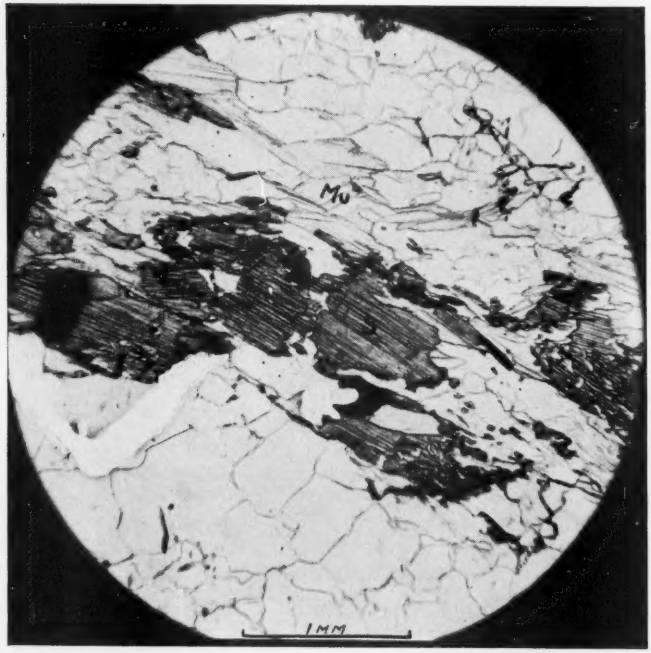


FIG. 8. Small spodumene grains in albite aplite. From the W. A. Ware farm, 2½ miles southwest of Kings Mountain, North Carolina. S = spodumene. Mu = muscovite.

Bands following the length of the broader pegmatites have been altered to albite aplite that seem to carry somewhat less spodumene. How much spodumene the albite aplite holds is not known, but probably not more than 2 or 3 per cent, if as much as that. I had hoped that, as at Tinton, South Dakota, some of the

fine-grained material would be rich in spodumene [22] but I have found none. In places the aplite is replaced in smaller or larger part by finely granular quartz and it too makes considerable bands. What part of the large pegmatites will prove to be of such materials is not known but a considerable quantity is to be expected.

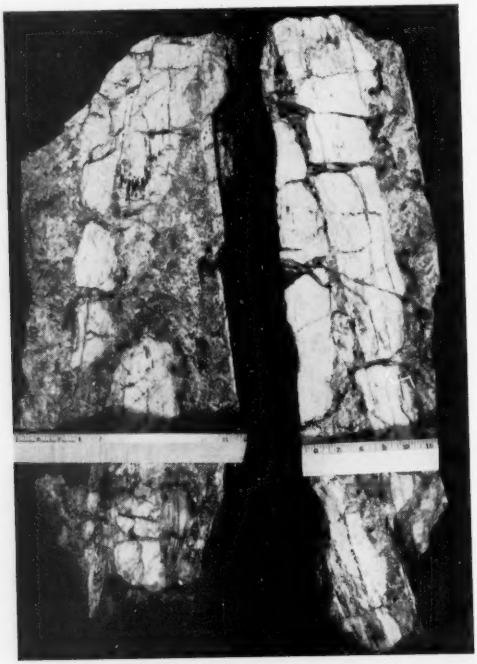


FIG. 9. Two views of spodumene crystals in the same specimen. From the W. A. Ware farm, 2½ miles southwest of Kings Mountain, North Carolina.

The paucity of muscovite in the spodumene pegmatites has been mentioned. This is so characteristic that in prospecting for spodumene it has been found unnecessary to examine pegmatites

[22] Hess, Frank L., and Bryan, Barnabas, Jr.: The pegmatites at Tinton, South Dakota. Bur. of Mines, Rept. of Investigations 3404: 1–19, 1938.

showing prominent mica. Spodumene pegmatites probably carry little muscovite because it has been replaced by spodumene. It is easy to write an equation showing how this replacement may have taken place, thus:

$$\underset{\text{Muscovite}}{2H_2O\cdot K_2O\cdot 3Al_2O_3\cdot 6SiO_2} + 3Li_2O + 6SiO_2 = \underset{\text{Spodumene}}{3(Li_2O\cdot Al_2O_3\cdot 4SiO_2)} + 2H_2O + 3K_2O.$$

Simple and probable as such a reaction seems, no partly replaced muscovite has been identified and wholly replaced mica probably would be unrecognizable. Partly or wholly replaced microcline is common (Figs. 6, 7). For the replacement of microcline the probable reaction seems to be:

$$\underset{\text{Microcline}}{K_2O\cdot Al_2O_3\cdot 6SiO_2} + Li_2O = \underset{\text{Spodumene}}{Li_2O\cdot Al_2O_3\cdot 4SiO_2} + 2SiO_2 + K_2O.$$

The K_2O freed from the microcline, together with SiO_2 taken from solution or quartz, might combine with muscovite taken into solution and make other microcline:

$$\underset{\text{Muscovite}}{2H_2O\cdot K_2O\cdot 3Al_2O_3\cdot 6SiO_2} + 2K_2O + 12SiO_2 = \underset{\text{Microcline}}{3(K_2O\cdot Al_2O_3\cdot 6SiO_2)} + 2H_2O.$$

thus possibly accounting for the late and fresh mineral shown in some spodumene pegmatites but not found here. It seems necessary that some such reaction as the first must take place because, as stated, nowhere do spodumene and much muscovite occur together.

The reactions for the other feldspars found in such pegmatites would be much the same as shown for the second reaction and should such reactions take place, they might account for part of the excess quartz found in most spodumene pegmatites.

Although original dark minerals are scarce, the pegmatite including the spodumene, is considerably stained by manganese oxide and by less iron oxide. The mobile oxides have been set free from the country rocks and carried into the pegmatites during the long period through which they have been exposed to weathering. Many cracks contain films of vivianite that may have been formed from either apatite or amblygonite and soluble iron minerals.

At a good many places, spots in the pegmatites are stained brownish yellow with a hint of green by nontronite, the hydrous iron silicate. Like the vivianite it is, of course, secondary. However, if the pegmatite can be mined below the weathered zone, it probably will be very low in iron.

ECONOMIC FEATURES.

Quality of Spodumene.—The spodumene is mostly very light yellowish and semi-opaque, although some crystals contain parts that are clear and colorless. Spectrochemical analyses on picked crystals show from 0.18 to 0.5 Fe_2O_3 (given as ferric oxide because all iron determinations for ceramic work are so reported). In 12 analyses of spodumene quoted by Dana [23] one gave only a trace, but the average of the other 11 was 1.06 per cent Fe_2O_3, with a low of 0.20 per cent and a high of 2.73 per cent. Dr. J. J. Kennedy [24] informs me that his analyses of spodumene from many parts of the world average about 0.5 per cent or more Fe_2O_3. From results at hand, it seems probable that the North Carolina spodumene, even where it is free from staining, carries a somewhat higher percentage of iron than does the remainder of the pegmatite.

The lithia content of the spodumene, as shown by three determinations made by Foster Fraas [25] was 6.28, 6.56, and 6.68 per cent, an average of 6.51 per cent Li_2O. Dana used an atomic weight of 7.0 for lithium and, by making allowance for the present accepted atomic weight of 6.94, in the 12 analyses that he cited the average lithia content was 6.55 per cent with a low of 4.45 and a high of 7.52 per cent, so that the Kings Mountain spodumene has an average content of lithia. Except locally, weathering seems to have caused little loss of lithia from the crystals in the outcropping rock.

Quantity of Spodumene Available.—As noted, some of the pegmatites are about a half mile long, are nowhere less than 20 feet

[23] Dana, J. D., and Dana, E. S.: System of Mineralogy, 6th edit. 1892, p. 367.

[24] Chief Chemist, Maywood Chemical Co., Maywood, N. J., most experienced American dealers in lithium.

[25] Personal communication.

wide, and reach a width of more than 100 feet, although some schist is probably enclosed in the wider sections. Others are much shorter and may have either a greater or a less width. The distribution of the spodumene in the pegmatites is much more even than the distribution of minerals in most metalliferous veins. What appeared to be representative samples have given between 15 and 20 per cent of spodumene concentrates, but there are much richer as well as much leaner parts.

Some of the hand-sorted spodumene mined by L. M. Williams on the W. A. Ware farm, carries as much as 4 per cent Li_2O so that it is more than half spodumene. Samples taken from a deposit of the Tennessee Mineral Products Corporation half a mile or more north of the W. A. Ware farm were about one-third spodumene.[26]

Experiments carried out by Norman and Gieseke[27] at the Eastern Experiment Station of the Bureau of Mines, College Park, Md., have shown that an excellent recovery of spodumene very free from other minerals can be made from the pegmatites.

It seems safe to estimate that if all the pegmatites were pieced together they would be more than equivalent to a single pegmatite 10 feet wide and extending the whole length of the belt, about 25 miles. Supposing a depth of 100 feet and a content of 15 per cent of spodumene having a specific gravity of 3.15 and a lithia content of 6.50 per cent, there would be available 1,650,000 short tons of spodumene containing more than 107,000 tons of lithia or 50,000 tons of lithium. The meagerness of the data used as a basis for this estimate is recognized and it is intended only as a more or less intelligent guess but is believed to be within the facts. The quantity of solutions that must have flowed through the rocks to carry this huge quantity of lithium and the other large quantity that undoubtedly flowed through without depositing was enormous and means that it was the exudation from a very large body of granite.

Possible Use.—So far as can now be seen, if there is to be much

[26] Norman, James, and Gieseke, E. W.: Beneficiation of spodumene rock by froth flotation. A. I. M. E. Tech. Pap. 1161: 2, 1940.

[27] Work cited.

of a future for spodumene, it will be in glass and ceramics. Although objection is made that for certain wares the iron content is too high to allow direct use of the mineral, it seems to be no higher and is probably somewhat lower than the iron in the aplite from Nelson Co., Virginia, now being used in large quantity for glass making, and in the lepidolite also now used. It would be possible to get rid of the iron by making lithium salts, but the salts would be much more expensive per unit of lithium and might not even otherwise be as good as the mineral, because the high alumina content of the spodumene (about 28 per cent) is a very desirable ingredient for glass.

As has been noted, it seems probable that the pegmatites as a whole contain less iron than the spodumene itself. It appears possible that the entire mass might be ground and, with the addition of soda or other necessary ingredients, be used as a batch for certain glasses.

Spodumene seems to have real advantages as an addition to ceramic mixtures.[28]

BUREAU OF MINES, EASTERN EXPERIMENT STATION,
COLLEGE PARK, MARYLAND,
May 14, 1940.

[28] Preston, E.: Lithium in glass and ceramics: Foote-Prints, 11: 1–15, 1938.

THE MOLYBDENITE DEPOSITS OF THE RENCONTRE EAST AREA, NEWFOUNDLAND.*

D. E. WHITE.

ABSTRACT.

Molybdenite occurs in the aplitic border phase of a Paleozoic batholith of granite and alaskite, which intrudes Cambrian quartzites and Ordovician (?) volcanics. In the Ackley City orebody, molybdenite is disseminated in altered aplite; a honeycomb of alteration was formed by the mineralizing solutions that penetrated the aplite without marked structural control. Silicification preceded muscovitization in space, but the two processes may be in part contemporaneous.

The mineralization is believed to be closely related genetically to the miarolitic aplite phase in which it occurs. A hypothesis is presented for the origin of the aplite, and its close relationship to the mineralization. The latter is believed to be due either to acid fluids—gases, or hydrothermal condensates, or to a combination of acid gases and alkaline liquids. Graton's hypothesis that ore-forming fluids are alkaline liquids fails to explain the relationships. The main period of mineralization is high-temperature with intermediate-temperature late phases. The Ackley City orebody is classified as a "complex aplite."

The Wylie Hill deposit is of intermediate-temperature and the Crow Cliff deposit is a pegmatitic type.

TABLE OF CONTENTS.

Introduction 968
Ackley Batholith 969
Granite 969
Alaskite 971
Aplite 972
Origin of the Aplite Phase 973
Depth of Intrusion 975
Molybdenite Deposits 976
Distribution 976
Ackley City Deposit 977
Wylie Hill 984
Mineralogy 985
Paragenesis 989
Origin 990
Classification 992
Acknowledgments 993
Bibliography 994

* Published with the permission of the Newfoundland Government Geologist.

INTRODUCTION.

RENCONTRE East is in the northern part of Fortune Bay, near the middle of the South Coast of Newfoundland (Fig. 1). An investigation made by the writer in the summers of 1937 and 1938 for the Geological Survey of Newfoundland showed that the rocks of the area range in age from pre-Cambrian to Carboniferous (?). Fossiliferous Cambrian rocks are succeeded by

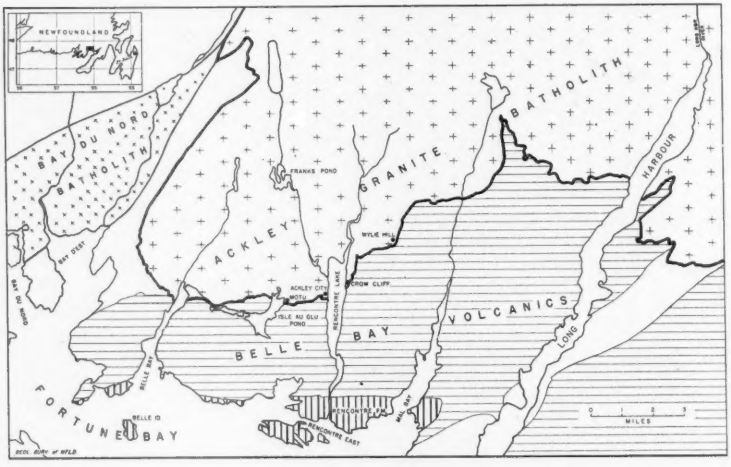


FIG. 1. Geologic map of the Rencontre East area.

three series of unfossiliferous volcanics and sedimentary rocks, with each series unconformable on the preceding.

A 13,000-foot succession of Ordovician (?) lavas and pyroclastics of terrestrial origin, the Belle Bay volcanics, is unconformable on the Cambrian rocks, and is the only stratigraphic unit of importance in relation to the molybdenite mineralization.

The general geology of the district is treated at some length in a forthcoming bulletin of the Geological Survey of Newfoundland.

ACKLEY BATHOLITH.

The batholith intrudes the northwest limb of a large syncline, the major structure of the Fortune Bay synclinorium. The invaded rocks are largely the Ordovician (?) Belle Bay volcanics, and to a lesser extent, tuffaceous slates conformably overlying the volcanics, and Cambrian quartzites. The mapped extent of the batholith is over 160 square miles, but this is probably less than half of the total. The long axis of the intrusion is oriented approximately northeast, parallel to the dominant regional structural trends. The dip of the contact, where it could be determined, is 25° to 45° outward from the batholith.

The topography of the batholith is of low relief, with elevations averaging about 750 feet, in contrast to the higher elevations and considerably greater local relief of the volcanics to the south.

The intrusion consists mainly of granite ("white granite") and alaskite ("red granite"), with the latter the more abundant, in the southern part of the batholith. These two phases are generally gradational, but sharp contacts and local cross-cutting relationships have been observed.

Basic and intermediate rocks are completely absent, although early phases of the differentiation series may be represented by the nearby Bay du Nord batholith.

Chilled contacts against the intruded volcanics and slates are rare, but have been observed.

Aplite is present in an irregular zone adjacent to the southern contact for a distance of about 2 miles west and 3 miles east of Rencontre Lake, and is most abundant in the large embayments into the intruded volcanics. The aplite is fine-grained, with a mineral and chemical composition similar to the medium-grained alaskite; the two phases are generally gradational over a variable distance, from a few inches to several feet. Cross-cutting dikes are rare; although aplite has been observed cutting alaskite (Fig. 2), no example of the reverse relationship was seen.

Granite.—The granite is rather coarse-grained, with pink orthoclase, and white albite-oligoclase, abundant quartz, and less than 10 per cent mafics. A representative modal analysis con-

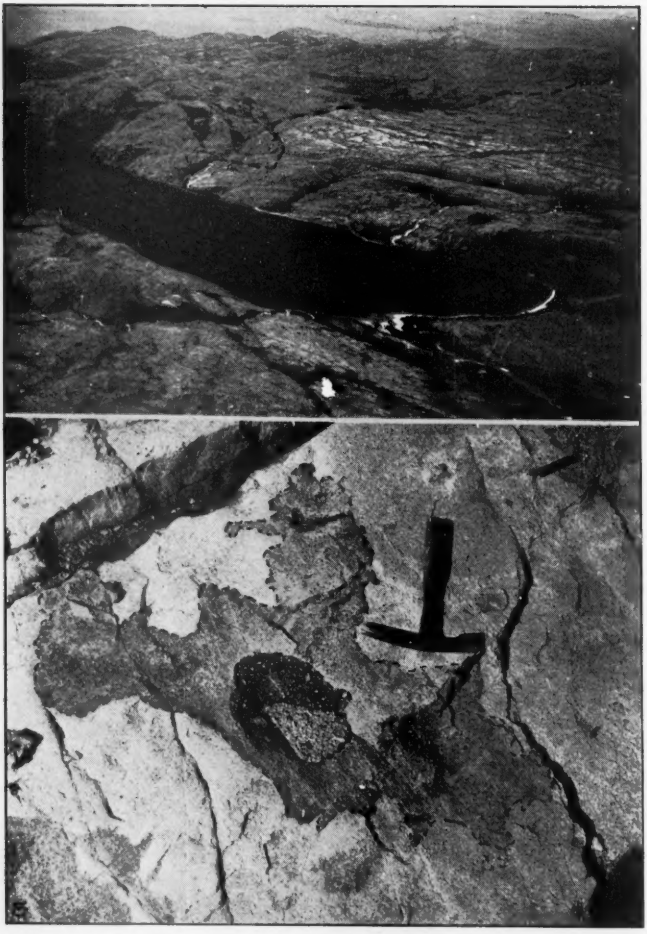


FIG. 2. Areal photograph looking southwest across Rencontre Lake; Ackley Batholith in foreground, with intruded volcanics in background. The Ackley City Mine is on the lake shore at the contact.

FIG. 3. Replacement of aplite, showing the zonal relationship, with silicification bordering muscovitization. Ackley City mine area.

sists of quartz 27 per cent, perthite 40 per cent, albite-oligoclase 25 per cent, 7 per cent biotite and chlorite, and 1 per cent miscellaneous. Every thin section shows some deuteric albitization of perthite, but the total change in composition cannot be determined.

Alaskite.—The alaskite is medium-grained, with a granitoid texture, and a pink color. A modal analysis consists of 39.5 per cent quartz, 33.8 per cent perthite, 25.9 per cent albite, and 0.9 per cent miscellaneous minerals, including biotite, chlorite, muscovite, magnetite, allanite, sphene, apatite, zircon. Both the perthite and albite are pink and indistinguishable megascopically from each other.

Rather extensive albitization of the alaskite, similar to that of the granite, is indicated by the thin sections.

TABLE I.

	Analysis.			Norm.			Mode.	
	1.	2.		1.	2.		1.	2.
SiO_2	79.07	77.33	Q	41.6	35.8	Qtz	39.5	37.0
Al_2O_3	11.65	12.56	Or	26.3	26.7	Perthite	33.7	31.8
Fe_2O_3	.36	.17	Ab	29.4	33.1	Albite	25.9	29.1
FeO	.28	.73	An	.6	2.2	Mafics	.9	2.1
MgO	.16	.12	C	.9	.4	Fluorite		.006
CaO	.09	.45	En	.4	.3			
Na_2O	3.45	3.90	Fs	.1	1.1		100.0	100.0
K_2O	4.46	4.54	Il	.2	.2			
H_2O^+	.25	.28	Mt	.5	.2			
				100.0	100.0			
H_2O^-	.06	.03		Plagioclase				
TiO_2	.11	.09		$Ab_{93}An_2$	$Ab_{91}An_8$			
	99.94	100.20						

1.—Medium-grained, slightly miarolitic red alaskite, from 65 feet north of the Ackley City wharf, and 10 feet west of Rencontre Lake. R. Folinsbee, analyst.

2.—Red miarolitic aplite, with some chlorite filling of the miarolites, from the east shore of Rencontre Lake, 125 feet north of the felsite contact. R. Folinsbee, analyst.

The silica content (Table I, no. 1), 79.7 per cent, is very high for an unsilicified rock. Vogt's composition-frequency curves of analyzed granitic rocks [20, p. 8] * shows the crest of the silica curve at

* Footnotes refer to bibliography at end of paper.

about 71 per cent. There is no microscopic evidence that the alaskite has been silicified, but deuteric silicification, if accomplished by large embaying replacements rather than by quartz veining, would be extremely difficult if not impossible, to distinguish from primary quartz.

Aplite.—The aplite phase of the batholith is of importance, because of the associated molybdenite. Most of the aplite is concentrated on the southern border of the batholith adjacent to the Belle Bay volcanics contact, in a 5-mile long zone that extends west to Isle au Glu Pond, and east to Wylie Hill. Aplite in this zone is intimately intermingled with alaskite, and is gradational into alaskite. Definitely cross-cutting dikes, characteristic of most aplite occurrences, are rare. The border zones of a few are gradational from aplite to alaskite, and consist of granophyric intergrowths.

The aplite is not a true chilled phase of the alaskite; although the general aplite zone bears a marginal relationship to the batholith, it is believed to be a slightly later phase that crystallized after the alaskite. The evidence is:

(1) In addition to the rare, small cross-cutting dikes that occur in the main border zone of the batholith, a few larger aplite dikes and irregular bodies are found within the batholith, unrelated to the contact (e.g., Long Harbor River).

(2) The gradation from alaskite to aplite does not seem to be, in general, a continuous grain-size gradation, but rather, crystals 1 to 2 millimeters in diameter increase gradually in proportion to larger crystals, 2 to 4 millimeters in diameter. Thus, the gradation zone consists of a porphyritic rock, with varying proportions of phenocrysts to groundmass.

(3) The aplite phase is commonly miarolitic, with the cavities or vugs generally less than ⅛ inch in diameter, but commonly up to ¼ inch, and rarely, over ½ inch. Careful observation of the aplite nearly always reveals miarolites, or relict miarolites, which have been filled with chlorite or quartz. Commonly, the euhedral quartz and feldspar crystals in the cavities are 2 or 3 times the size of the g[illegible]ins of the main aplite.

The texture is hypidiomorphic, with both feldspars commonly euhedral or subhedral. Most of the quartz is contemporaneous with the feldspar, but some is later. Miarolitic cavities lined with euhedral crystals were probably most abundant at the end of the primary crystallization stage, with later filling of some by deuteric deposition of quartz. Late quartz in the cavities has the crystal habit of α-quartz, probably formed below 600° C. at the pressure existing during crystallization of the magma. Granophyric intergrowths of quartz and orthoclase, or quartz and albite are commonly found in aplite, but almost never in alaskite. In contrast, perthitic texture is common in alaskite but is rare in aplite.

(4) Zoned textures immediately at the contact of aplite with the Belle Bay felsites are common and are also found surrounding felsite inclusions. Several specimens show large euhedral quartz and feldspar crystals that have grown outward from inclusions (Fig. 4). Another specimen shows zoned intergrowths away from the contact, with a narrow zone of albitization or primary albite adjacent to the contact. These zonings are strong indications that the aplite is not an early chilled phase of the alaskite.

Analysis No. 2, Table I, is representative of the primary aplite phase, which has undergone slight deuteric action, in the partial filling of miarolitic cavities by chlorite. The modal analysis, similar to that of the alaskite, shows perthitic feldspar in excess of albite, whereas the normative calculations indicate a greater abundance of albite.

Although the aplite is later than alaskite, it is slightly lower in SiO_2, and in the ratio of orthoclase to plagioclase, and is higher in the mafics than the alaskite. Deuteric processes are believed to account for the greater amounts of albite and of chlorite in the aplite.

Origin of the Aplite Phase.—One possible interpretation for the origin of the aplite is as follows:

During the last stages of crystallization of the magma, the residual liquid became much enriched in the volatile constituents, with a resultant rapid increase in the internal vapor pressure.[18] When the pressure of the overlying load was finally equalled by

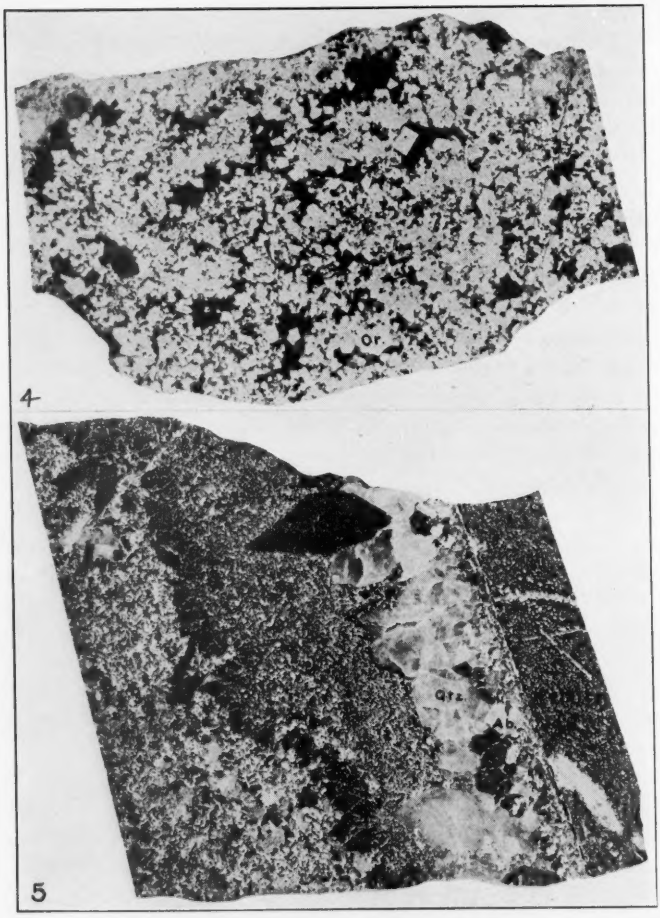


FIG. 4. Euhedral habit of the feldspar of the potash-rich aplite. Etched hand specimen, reflected light, with orthoclase (light gray) and quartz (black). Both albite and molybdenite are present, but albite is indistinguishable from orthoclase, and molybdenite from quartz. × 2.

FIG. 5. Euhedral crystals and zoned intergrowths about an inclusion immediately adjacent to the contact of aplite with felsitic rhyolite. Etched hand specimen, with albite (white), quartz (light gray), orthoclase (dark gray). × 1¾.

the internal vapor pressure, a boiling off, or a distillation, of the volatile constituents was the natural result.

Previous to this distillation, granite and then alaskite had been the phases that crystallized. None of the granite, and only minor alaskite contains miarolitic cavities, but as soon as the boiling process became important, the *rate* of crystallization increased, because removal of heat by escaping super-heated volatiles was *added* to normal removal of heat by conduction through the wall rock. The increased rate of precipitation resulted in a finer-grained rock—an aplite with a miarolitic texture due to entrapped gases.

The proposed origin for the aplite is similar to that suggested by Andersen[1] for some aplite dikes.

Depth of Intrusion.—Vogt [20, pp. 172–177] has suggested the following criteria as distinguishing between shallow-seated and deep-seated intrusions. The starred criteria have been suggested by Buddington.[5]

Shallow-Seated	Deep-Seated
1. Aplite as late differentiate.	1. Pegmatite as late differentiate.
2. Granophyric intergrowths.	2. Myrmekitic intergrowth of plagioclase and vermicular quartz, replacing potash feldspar.
3. Soda pyroxenes, and no muscovite, (due to escape of volatiles).	3. Micas, hornblende (volatiles reacting in late stages).
4. * Quartz-albite granites.	4. * Albitization not pronounced.
5. Miarolitic texture.	5. No miarolitic texture.

If these criteria are valid, the Ackley batholith is a relatively shallow-seated intrusion because of the great preponderance of aplite over pegmatite; the rare granophyric texture and the absence of myrmekite; the albitization of all phases of the intrusion; and the presence of miarolitic texture, particularly in the late phases.

The femic constituents are mica and hornblende, which never exceed 10 per cent of the total rock, and are negligible in the alaskite and aplite phases. The presence of biotite and hornblende suggests that the Ackley batholith was intruded near the base of the shallow-seated zone. The absolute depth of this zone cannot be determined.

MOLYBDENITE DEPOSITS.

Distribution.—Nearly all of the known molybdenite of the Rencontre East area is restricted to the southern border of the Ackley batholith, at or near its contact with the Belle Bay volcanics. This portion of the border zone is about 5 miles long, and extends from Wylie Hill on the east to Isle au Glu on the west. Molybdenite is concentrated at four localities: Wylie Hill, at the eastern end; Crow Cliff and the Ackley Mine, on the shores of Rencontre Lake, near the center; and Motu, near the western end of the zone (Fig. 1). The central mineralized areas adjacent to Rencontre Lake are separated from Wylie Hill and Crow Cliff by unmineralized and sparsely mineralized rock.

The distribution of the molybdenite shows a close correlation with the distribution of the aplitic phase of the Ackley batholith; the richest mineralization is always in aplite, but alaskite also contains molybdenite where it is intimately associated with mineralized aplite.

Most of the aplite is concentrated in three broad embayments of the batholith into the volcanics, namely, Wylie Hill, Rencontre Lake, and Motu areas, each of which contains within itself a smaller zone of mineralization adjacent to the contact with the Belle Bay volcanics, which form the hanging walls of each of the principal deposits. However, many aplite areas of various sizes contain no molybdenite, particularly where they are unrelated to the volcanics contact.

The Ackley City deposit is the most promising and most of the development work has been done on it. The Wylie Hill deposit is much larger than the Ackley City deposit, but is of a considerably lower grade. The Crow Cliff and Motu deposits show erratic distribution of molybdenite, but small tonnages of commercial ore may be present. Crow Cliff is a pegmatitic type of mineralization; Motu is like the Ackley City type.

The only mineralized area in the Ackley batholith that bears no relation to the contact lies southeast of Franks Pond, over 3 miles north of the contact. However, this deposit differs from that of the other areas because molybdenite is associated with quartz

veins in granite. Two mining claims carry rather widely spaced veins. Each vein is several inches wide, and is sparsely mineralized with molybdenite; those of the northern claim are in general oriented N. 72° E., and those of the southern claim almost at right angles, N. 17° W.

Ackley City Deposit.—Each area has certain distinctive characteristics; one type of ore is represented by the Ackley City deposit, in which molybdenite is disseminated in altered aplite on the west shore of Rencontre Lake, immediately at the contact with the Belle Bay volcanics. The hanging wall contact strikes about N. 75° E., and dips 45° to 50° S.

The orebody as now known is a lens-shaped body about 140 feet long that plunges eastward toward Rencontre Lake. The width is about 40 feet; its south boundary is the hanging wall volcanics, and its north boundary is an assay ore.

To the end of 1938, development work included 1,000 feet of surface trenching, a 60-foot shaft, and 650 feet of underground work.

The ore zone consists of numerous patches of a greisen-like type of alteration, with quartz, muscovite, molybdenite, fluorite, and chlorite distributed through aplite. Orthoclase inversion to microcline is pronounced, particularly on the border of the ore zone. Biotite, magnetite, hematite, pyrite, chalcopyrite, pyrrhotite, and sphalerite are minor associations.

The mineralizing solutions have permeated the aplite to form a "honeycomb" of alteration patches, generally without marked structural control, except during the late alteration stages. The resultant orebody is mineralized throughout, but *in detail* the alteration and molybdenite content are irregularly distributed.

Zoning of the quartz and muscovite replacements on a small scale (Fig. 3), and possibly on a large scale, is indicated, with the silicification in advance of the muscovitization in space, but possible in part contemporaneous in time. Immediately west of the orebody, which is also, in general, the western limit of muscovitization, the adjacent alaskite shows some silicification, suggesting a large-scale zoning of silicification and muscovitization. However, definite zonal relationships are commonly absent, indicating

that the muscovitizing solutions did not always follow the channels of the earlier silicification; thus each type of alteration is in places independent of the other.

Molybdenite is apparently more abundant in the replacement quartz that accompanies areas of muscovitization, but it is also distributed throughout the muscovite rock. Also, some of the highest grade ore consists of disseminated molybdenite in relatively unaltered aplite.

During the silicification of the aplite, the potash feldspar was in some instances selectively replaced in preference to albite. A detailed study would be necessary to prove general selective replacement.

The chemical analysis of the muscovitized rock (No. 3, Table II) shows the pronounced changes that took place during the alteration.

A comparison of the accompanying mineralogical changes emphasizes the degree of alteration the aplite has undergone.

Is the muscovitized rock a replacement of unaltered aplite, or of silicified aplite? During silicification, which may be earlier, at least in part, orthoclase is selectively replaced in preference to albite; however, the muscovitized rock consists essentially of muscovite and orthoclase, with minor amounts of quartz and albite. The chemical additions and subtractions of all the common oxides are exactly opposite in the two types of alteration, as shown in Table II, except for Na_2O, which decreases gradually in relation to K_2O during silicification, in contrast to a pronounced decrease during muscovitization.

The abundance of orthoclase and the rarity of quartz in the muscovitized aplite suggests the alteration of a *non-silicified* aplite. How can this be reconciled with the zonal relationship that shows silicification in advance of muscovitization in space, and possibly in time?

The possibilities are: (1) The muscovite-orthoclase rock is a direct replacement of unsilicified aplite. In this case, the orthoclase is probably primary, and the muscovite has selectively replaced quartz and albite. Where such a rock is zoned by silicification, the SiO_2 which was leached from the muscovite zone may

TABLE II.

	Chemical analyses.		Per cent changes during muscovitization of aplite.	Changes during silicification of aplite.
	No. 2.	No. 3.		
SiO_2	77.33	48.93	−37%	+
Al_2O_3	12.56	26.49	+111	−
Fe_2O_3	.17	2.68	+1470	−
FeO	.73	3.64	+400	−
MgO	.12	.49	+310	−
CaO	.45	1.46	+220	−
Na_2O	3.90	1.36	−65	slightly −
K_2O	4.54	9.31	+105	−
H_2O^+	.28	3.09	+1000	−
H_2O^-	.03	.04	+33	
TiO_2	.09	.16	+78	
P_2O_5		.01		
F		1.38		
S		1.27		
MnO		.42		
Mo		.19		
	100.20	100.92		
		1.01 Less O_2		
Total	100.20	99.91		

	Modal analysis.	
	No. 2.	No. 3.
Quartz	37.0	1.7
Or + Perthite	31.8	20.2
Albite	29.1	4.5
Muscovite		64.0
Fluorite	.006	1.7
Chlorite	2.1	5.9
Magnetite		2.0
	100.0	100.0

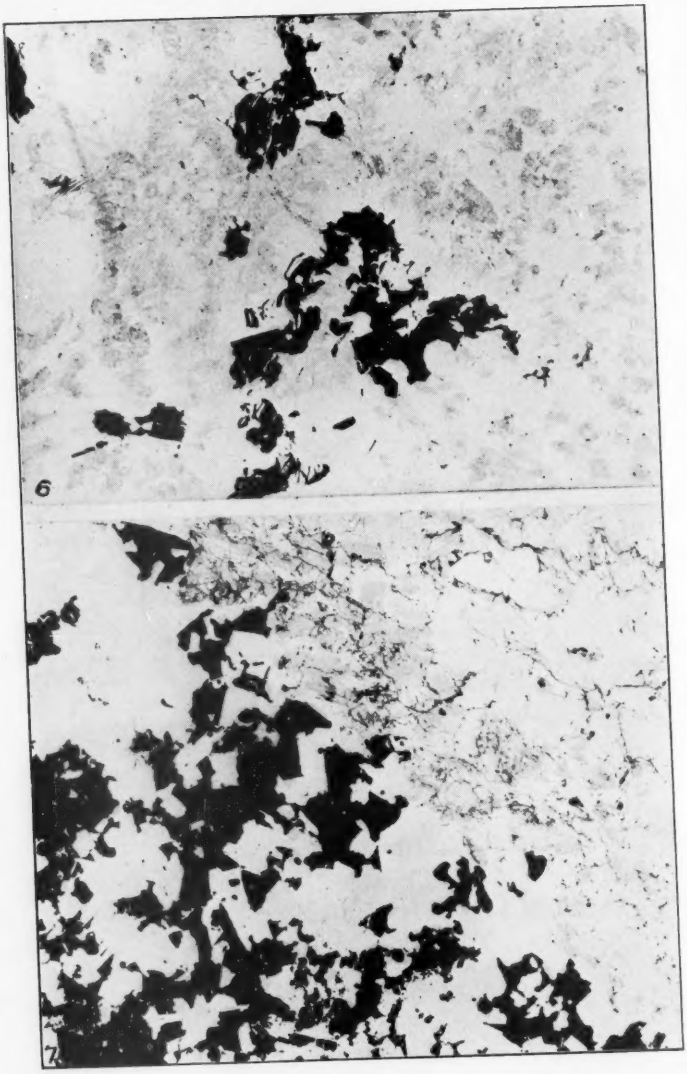
No. 2—Normal aplite, 125 feet north of the batholith contact, east shore Rencontre Lake. R. Folinsbee, analyst (No. 2).

No. 3—Muscovitized aplite, east crosscut, 60 foot level, Ackley City Mine. R. B. Ellestad, analyst (No. 378f).

have selectively replaced orthoclase in the outer silicification zone. A quartz-muscovite rock, which is less common than the muscovite-orthoclase rock, may thus be a final encroachment of muscovite into the silicified zone. The microscopic evidence would seem to favor this explanation of the relationships.

(2) The second possibility is a replacement by muscovite *and orthoclase* of previously silicified aplite. Orthoclasization has been described by Schwartz[15] and Gillson,[9] but is apparently a much rarer process than albitization. No direct evidence for orthoclasization has been observed in the thin sections.

Molybdenite mineralization was either contemporaneous with the late stages of silicification and muscovitization, or closely fol-



FIGS. 6–7.

lowed them. Locally, molybdenite-depositing solutions were diverted from channels of previous alteration; disseminated molybdenite was then deposited in fresh-looking, relatively unaltered aplite (Fig. 6).

TABLE III.

	Chemical analyses.			Modal analyses.	
	No. 2.	No. 4.		No. 2.	No. 4.
SiO_2	77.33	75.05	Quartz	37.0	37.5
Al_2O_3	12.56	11.13	K-feldspar	31.8	40.7
Fe_2O_3	.17	.26	Albite	29.1	13.5
FeO	.73	.57	Muscovite		2.8
MgO	.12	.14	Molybdenite		4.9
CaO	.45	.54	Chlorite	2.1	.5
Na_2O	3.90	1.92	Fluorite	.006	.1
K_2O	4.54	6.43			
H_2O^+	.28	.21		100.0	100.0
H_2O^-	.03	.01			
TiO_2	.09	.13			
P_2O_5		.02			
F		.35			
S		1.49			
MnO		.04			
Mo		1.99			
		100.28			
		.21 Less O_2			
Total	100.20	100.07			

No. 2—Normal aplite, 125 feet north of batholith contact, east shore Rencontre Lake. R. Folinsbee, analyst.

No. 4—Molybdenite-aplite. Dump specimen, Ackley City Mine. R. B. Ellestad, analyst.

Analysis No. 4, Table III, is of a high-grade molybdenite-bearing aplite. The molybdenite is associated with a little muscovite, but the most pronounced difference between this aplite and the normal aplite, No. 2, Table III, disregarding MoS_2, is in the ratio

FIG. 6. Molybdenite disseminated in relatively unaltered potash-rich aplite, with the molybdenite intergrown with some muscovite. Thin-section. × 13.

FIG. 7. Pyrite intergrowth; border of a nodular replacement of normal aplite. The pyrite in part selectively replaces the primary quartz of the aplite. A muscovite-rich iron-stained zone borders the pyrite. Thin-section. × 7.

of the feldspars. Although the total feldspar is about equal in both rocks, orthoclase makes up 53 per cent of the total feldspar of specimen No. 2, in contrast to 73.8 per cent of the total feldspar of the molybdenite-bearing aplite, No. 4. The pronounced euhedral habit of the orthoclase and microcline is notable, and is illustrated in Fig. 4.

The unusual nature of this orthoclase-rich aplite is shown by reference to Vogt's frequency curves [20, p. 45] of orthoclase-albite ratios of analyzed aplites. Vogt has no record of an aplite with over 60 per cent of the total feldspar consisting of orthoclase.

An orthoclase percentage of 73.8 per cent is very near Vogt's maximum recorded limit of the silica-rich granites; [20, p. 23] no granite with as high a silica content, in addition to as high an orthoclase percentage is recorded in his tables.

The nearest chemical equivalent to No. 4 among the intrusive igneous rocks is the pegmatite group, [20, p. 69] with an orthoclase maximum of approximately 80 per cent of the total feldspar.

Of the extrusive rocks, the pitchstones are the only rocks of similar composition to the orthoclase-rich aplite. Bowen's K_2O and Na_2O curves plotted against SiO_2 percentage,[4] have a maximum SiO_2 content of about 77 per cent. The 76 per cent SiO_2 is nearly comparable to this aplite, with K_2O at about 6 1/3 per cent and Na_2O at about 1 2/3 per cent of the total composition.

The author was not aware of this potash problem in the field; both aplites are strikingly similar in appearance, because both feldspars are red, and are indistinguishable megascopically. The only locality where the K_2O-rich aplite has been recognized by thin sections is the Ackley City Mine area.

Three possible explanations of the relationships of the two aplites are:

1. Orthoclase replacement, to form the K_2O-rich aplite. Microscopic investigation has failed to reveal adequate evidence for such a replacement, either of albite, or of quartz that had previously replaced albite. Several orthoclase crystals contain albite centers, which may be relicts of larger albite crystals, but the boundaries between feldspar grains are ordinarily rectilinear,

without definite embayments or veins of orthoclase in albite. Also, the strikingly euhedral habit of the orthoclase does not seem to be in harmony with a replacement origin.

2. The "normal" aplite may be an albitized aplite, originally potash-rich. Evidence for albitization is more abundant in specimen No. 2, than of orthoclasization in No. 4. The best evidence against this possibility is based upon the composition-frequency curves previously mentioned, which indicate that No. 2 is a completely normal aplite, and that No. 4 is very abnormal in its abundance of K_2O.

3. Both the normal aplite and the potash-rich aplite may be essentially primary, with the latter a late K_2O-rich differentiate of the former. This view is most acceptable at the present time, principally because the difficulties associated with the other possible modes of origin are avoided. No field evidence whatsoever was observed in support of this theory; however, the problem was not realized until after the completion of the field season. This hypothesis is based on Terzaghi's belief [17] that some of the potash-rich rocks are explainable in the following manner: "Toward the end of the crystallization process, when water is present in relatively large quantities in the liquid phase, potash and plagioclase feldspar may be in equilibrium with a liquid having a higher Or: Ab + An ratio than if the melt is 'dry.'"

Separation of such a potash-rich liquid from a magma that is rapidly crystallizing as normal aplite involves certain difficulties. The rare occurrence of potash-rich aplites may mean, therefore, that the necessary combination of chemical, structural and dynamic factors necessary for their formation are rarely attained.

Fracturing, with the formation of the earliest prominent joint systems, is believed to have occurred during the late stages of the quartz-muscovite-molybdenite mineralization; commonly a N. 80° E., 80° N. system near and in the Ackley City deposit carries a replacement zone 1 to 2 inches wide, of silicification and muscovitization and, rarely, iron oxides and molybdenite. A N. 10 to 20° W., 80° E. system in places shows the same alteration, but more commonly is unaltered or carries quartz veinlets. Much

less prominent systems, N. 40° E., and N. 70° W., have been observed with the same alteration association.

The common joints belong to the N. 70° to 90° E. and the N. 10 to N. 20° E. systems, which are generally predominant both in the batholith and in the volcanics, from Long Harbor to Belle Bay, and particularly from Wylie Hill to Isle au Glu Pond. Approximately 300 joint-orientation readings were taken in this part of the area, in determining the major systems.

Others are locally important or predominant. Minor frequency maxima at N. 30° E., and N. 55° E. have been noted; a very even distribution of joints occurs between N. 30° W. and N. 80° W., with no prominent maxima.

Wylie Hill.—The Wylie Hill deposits consist of two large mineralized areas, each approximately 1200 feet long by 400 feet wide. Both areas are adjacent to the contact of the Ackley batholith with the Belle Bay volcanics, with the latter forming the hanging wall of the orebodies. Both deposits are predominantly aplite and, in general, have a definite boundary against non-mineralized alaskite, but minor amounts of alaskite have been altered, and some aplite is present outside of the alteration boundaries.

The principal characteristics of the two deposits are: (1) The joint systems are closely spaced, averaging between 2 to 6 inches between joints, in contrast to a spacing of several feet or more between joints in the alaskite. The predominating joint systems in the Wylie Hill area, N. 10° to 20° W., 75° W.; and N. 75° to 85° E., 65° N., belong to the same predominant systems that are represented both in the batholith and in the adjacent lavas, extending at least from Wylie Hill to Isle au Glu. A minor joint system is prominent in the 15-foot pit on the shore of Wylie Pond. These joints control the quartz-pyrite-molybdenite mineralization.

(2) Abundant iron-staining of the mineralized rock is due to surface weathering of disseminated pyrite in the aplite. The pyrite is in places evenly disseminated, but is more commonly present in veinlets and in nodule concentrations (Fig. 7), which weather characteristically.

(3) The aplite is commonly porphyritic; a fresh surface is light gray or light pink in contrast to the red color of most of the

aplite of the area; such a color change is apparently a constant association of pyritization. Iron, present in the red feldspars may have been leached out during mineralization to form at least a part of the pyrite.

(4) Molybdenite is present generally as thin quartz-pyrite-molybdenite joint-coatings. Muscovite and disseminated molybdenite are comparatively rare. Although joint blocks commonly appear to be very rich ore, this impression is deceptive, because the molybdenite is present only as joint coatings.

(5) The best ore is adjacent to the contact with the volcanics, but the relationship is not as pronounced as in the Ackley City deposit.

The molybdenite content of most of the two areas is considerably below ore grade, but several small patches along the eastern contact of the east area are of ore grade.

Mineralogy.—Molybdenite is the principal and only economic mineral found in the Rencontre East area. It is present in a variety of associations, one of which is a dissemination in silicified aplite, muscovitized aplite, and in relatively unaltered aplite (Fig. 6). It commonly selectively replaces quartz but also is found with muscovite in parallel intergrowth. No conclusive evidence was seen to indicate whether the intergrowth is due to contemporaneous deposition, or to replacement of one component by the other. Fluorite, in variable amounts, is an almost constant association of molybdenite. Where the molybdenite is disseminated in aplite, it is generally found as crystals and rosettes up to ½ inch in diameter, but larger crystals are not unusual, particularly in highly silicified aplite.

One specimen shows molybdenite with a pearly carbonate, which it is probably replacing, in a small silicified area. However, the relation of the carbonate to the paragenetic sequence is not known, except that it is later than the silicification.

An association with pegmatitic quartz and feldspar is rare, but a quartz-vein association, with variable amounts of biotite, chlorite, fluorite, magnetite, and pyrite is not uncommon.

The Franks Pond mineralization consists of quartz-molybdenite veins, rarely over a few inches wide; quartz-pyrite-molyb-

denite veinlets and joint-plane coatings are typical of Wylie Hill, and quartz-molybdenite-chalcopyrite-fluorite-carbonate veinlets of a prospect on the east shore of Belle Island, in Fortune Bay.

Pyrite is minor in the Ackley City deposit, but may be as abundant as molybdenite in the highly muscovitized rock, and is the predominating sulphide in the Wylie Hill deposit, where it is found as disseminated, irregularly spaced nodules in aplite, as even disseminations, and as veinlets and joint-coatings with quartz and molybdenite. Each nodule is a selective intergrowth replacement (Fig. 7) of the quartz of the aplite, all of the pyrite in a single nodule generally having the same orientation. Pyrite is earlier than molybdenite at Wylie Hill and is probably earlier in the Ackley City deposit, although the evidence for the latter is not conclusive.

The feldspars of the pyritized aplite at Wylie Hill and locally at Motu have been leached of their characteristic red color; practically no leaching of the color has been observed where pyrite is not present. A possible explanation is that iron, present in the feldspars either as disseminated grains of oxide, or as an isomorphous constituent of the feldspar molecule—such as iron orthoclase,[8] $KFeSi_3O_8$—has been leached from the feldspars to combine with sulphur supplied by the mineralizing solutions.

Magnetite is found in minor amounts in the Ackley City deposit; most of it is associated with muscovite and chlorite, but small octahedral crystals have been observed in open cavities with quartz. Magnetite may be at least in part later than molybdenite.

Hematite is very minor, but is found disseminated (as replacements of magnetite?), and as euhedral crystals in quartz-hematite-muscovite replacement veins.

Pyrrhotite is also sparse, and has been identified only in the silicified alaskite zone west of the boundary of the Ackley City ore body.

Sphalerite has been identified in the Ackley City deposit as disseminations in one specimen of muscovitized aplite (dark sphalerite) and in several late veinlets with quartz, carbonate, and galena (light sphalerite).

Chalcopyrite is also rare; exsolution blebs in dark sphalerite

grains disseminated in muscovitized aplite, and a small octahedral crystal in an open cavity have been observed in the Ackley City deposit; it is also found in veinlets with quartz and molybdenite at Belle Island.

Galena is a minor constituent, and has been identified only in late veinlets with quartz, sphalerite, and carbonate.

Quartz is the most abundant primary constituent of the aplite, and as a replacement of aplite is an abundant product of the earliest stage of mineralization. Late-stage quartz veinlets are unimportant in the Ackley City deposit, but carry the molybdenite in the Wylie Hill deposit and the Franks Pond area.

Orthoclase is an abundant primary constituent of the aplite, and also may be in part a replacement mineral of the early stages of mineralization. A few narrow quartz-orthoclase-fluorite-pyrite veinlets that cut previously muscovitized aplite are probably later than molybdenite.

Microcline in the Ackley batholith is mostly restricted to the Ackley City deposit, where most of the orthoclase in the silicified alaskite west of the ore body has inverted to microcline; some microcline is present throughout the ore body, but much of the orthoclase has not inverted. The relative time of inversion in the paragenetic sequence is unknown; it may be an equilibrium response [16] to the submagmatic temperature and the abundant volatile constituents present during the early stages of the mineralization, or the inversion may be due to the stress conditions that produced the abundant joints, and local minor brecciation, later than most of the mineralization.

Albite is an abundant primary mineral and is also in part a deuteric replacement mineral throughout the Ackley batholith.

Muscovite is the most characteristic associate of molybdenite, particularly of the Ackley City type of ore, but it is also present in minor amounts at Wylie Hill, and its equivalent, sericite, is important at Belle Island. Megascopically, it is light to dark green, or yellow-green in color. The grain size varies from fine-grained aggregates to crystals ¼ inch in diameter. Most of it is a true muscovite, but some, particularly that replacing albite, is more correctly termed sericite. The composition of the pure

mineral is unknown, but the analysis of the muscovitized rock (No. 3, Table II) when compared with the modal analysis, indicates the probable presence of fluorine, iron, and possibly manganese.

Most of the muscovite appears to be earlier than molybdenite, but some may be contemporaneous. The strong euhedral habit of both minerals results in a singular absence of cross-cutting relationships.

Biotite is minor in the ore deposits; it is most commonly present as large euhedral or subhedral plates, oriented perpendicular to the walls of quartz-molybdenite-fluorite-chlorite-bagnetite veins. Most of these plates have altered to a green, pleochroic chlorite. Biotite is earlier than molybdenite. A fine-grained greenish biotite, with fluorite, is locally disseminated in aplite. This biotite is an unusual type, with a fairly high axial angle for that mineral, $25° \pm$, and a light greenish pleochroism resembling much of the chlorite. A moderately high birefringence, however, is indicative of biotite.

Fluorite is a common associate of muscovite and molybdenite, and is present in variable amounts as a light greenish variety. Purple fluorite is minor, and is probably late. Small euhedral crystals of fluorite, possessing a cubic habit modified by both dodecahedron and octahedron, were found in the same cavity which yielded an octahedral chalcopyrite crystal, as well as small doubly terminated quartz crystals.

Carbonate is sparse, and is present mostly in the late veinlets. Both calcite and dolomite are present; small crystals of the latter have been observed in cavities.

Chlorite is nearly always present in minor amounts in the Ackley City type of ore as an associate of muscovite, or as a replacement of biotite. A large part of the chlorite is a clear green, biaxial negative, pleochroic variety, with a birefringence of $.008 \pm$, and may be a relatively high temperature variety. Later (?) chlorite is of a "muddy" green color, with low birefringence, and anomalous interference colors. Some of the chlorite is earlier than a part of the molybdenite.

Sanidine is an unusual mineral to find in an ore association,

but it has been identified in two thin sections; one is a quartz-hematite-sanidine replacement of aplite, 700 feet northwest of the Ackley City mine shaft; the second thin section is of a quartz-muscovite-sanidine replacement vein 1000 feet ± west of the mine shaft, and is the same type of replacement vein that is common in the Ackley City deposit, at least in part later than the quartz-muscovite-molybdenite alteration patches. The identification of this mineral as sanidine is based principally on its very small axial angle. Its birefringence is also somewhat less than orthoclase.

The presence of sanidine in replacement veins of this type is unusual. Spencer [16] believes it is the highest temperature potash-feldspar, with an equilibrium temperature of over 900° C. It may be present in the Ackley City deposit due to an inequilibrium effect, possibly a result of pneumatolitic gas action.

Paragenesis.—During the early high temperature stages of formation of the Ackley City deposit, solutions that deposited quartz, muscovite, and molybdenite with minor chlorite, fluorite, and magnetite intimately permeated the aplite, without marked structural control. The primary fracture system is believed to have formed after these earliest stages; the replacement quartz-muscovite-sanidine veinlets, with minor molybdenite, in general did not control the main mineralization, and therefore are believed to represent the last part of the silicification-muscovitization stage.

At least one period of fracturing, with minor brecciation, followed the earlier joints; the intensity distribution of the later joints seems to be correlated with the degree of the earlier mineralization, but this is probably a localization due to the relative incompetence of the muscovitized rock; faulting and displacements along the joints are negligible.

Late mineralization is negligible; a specimen of molybdenite replacing carbonate may be relatively late. The common sulphides, except pyrite, are rare, and are present only as late-stage, joint-filling veinlets.

In the Wylie Hill orebody the dissemination of pyrite may have been earlier than the jointing, or may have been controlled by joint-plane mineralization of quartz-pyrite-molybdenite, with molybdenite after pyrite.

Origin.—Two particularly unusual features are shown by these ores: 1. Very intimate relationship to the aplite. These deposits seem to be unusually clear cases of genetic relationship of a mineralization to the associated igneous rocks. They are pertinent evidence in the refutation of a contrasting theory of ore genesis, which denies any genetic relationship of a mineralization to the associated igneous rocks; this view has recently been supported by Holmes [11] and by Billingsley and Locke.[2]

2. The second unique feature of the ores is their presence so far down the flanks of a large batholith, in contrast to the usual ore concentration in or near cupolas of small masses.

The following picture of the sequence of events from the formation of the aplite through the stages of mineralization is a possible explanation for these unusual features:

A late-stage magma, mainly concentrated in broad embayments of the batholith, crystallized as a fine-grained aplitic rock because of its relatively rapid rate of crystallization, which was due to the removal of magma heat by escaping volatiles, plus removal of heat through the wall rock.

After complete crystallization at the present level of erosion, gases continued to rise from the magma crystallizing at a greater depth. The natural channelways of these rising gases were the crests of the broad, embaying, inverted troughs of aplite, dipping outward from the batholith.

Such a structural control is similar to the control proposed by Emmons,[7] for the porphyry copper type of mineralization.

Alteration and mineralization were accomplished either by these rising gases or by hydrothermal solutions that resulted from the condensation of the gases.

The relative restriction of the mineralization to aplite, where alaskite was also present nearby, may possibly be due to a slightly greater permeability of the aplite; this would certainly be the case if the aplite actually crystallized while gases were streaming through, as has been proposed. In contrast, the alaskite, if crystallized without streaming gases, might have as great a porosity as the aplite, but would not be as permeable.

The adjacent Belle Bay lavas were relatively impervious to the

mineralizing solutions. The only molybdenite that has been observed in the volcanics is in two small quartz-hematite veins, 1 foot wide and 3 inches wide, respectively, several hundred feet south of the contact at Wylie Hill.

The mineralizing fluids consisted at least in part of acid gases,* or of liquids condensed from gases. The genetic relationship of the mineralization to an aplite with a miarolitic texture, plus the abundance of fluorine in the ore is of great significance. The texture of the aplite proves the presence of a gas phase during the late stages of crystallization of the magma. The late residual magmatic solutions that remained after appreciable separation of a gas phase would have been deficient in fluorine, which readily forms volatile compounds, and would therefore have been incapable of supplying the fluorine in the ore. The presence of a gas phase at some time during the separation of the ore-forming fluids from the magma is in accord with the views of Bowen,[3] based upon physico-chemical experimentation, and the evidence from the field relationships, reviewed by Schmedeman.[14]

Graton [10a] has recently presented a review of the evidence pertaining to the nature of the ore-forming fluid, and attempts to prove that alkaline liquids have formed the great majority of ore deposits. Some of his arguments have weaknesses; his best support is found in the "potash problem." Schwartz [15] has shown that in the hydrothermal alteration of igneous rocks, the average igneous rock shows a gain in potash, but no effective means of transportation of potash in a vapor phase is known at the present time. However, the vapor pressures of all possible potassium compounds existent at magmatic temperatures and pressures are not yet known. Graton's proposal that the motive power which elevates the liquids is the hydrostatic pressure of the overlying rock column fails to recognize the pronounced contraction that the igneous body is undergoing during the very late stages of its consolidation. The strength of rock units, from a large to a microscopic scale, and their porosities are factors that prevent the full

* The term "acid gases" is an incorrect use of "acid" in the chemical sense. It is meant to include all of the gases and gas mixtures which give acid reactions upon condensation to liquids.

development of a hydrostatic pressure equal to the overlying rock column on an igneous mass that is contracting.

A mixture of gas bubbles and liquid with the gas phase transporting the fluorides and providing the motive power, and the liquid phase transporting the potash may be necessary to explain these molybdenite deposits.

Classification.—The Ackley City mineralization is a high temperature type, with minor intermediate temperature effects during the late stages. Most of the minerals of this assemblage are not restricted to a single temperature group, but muscovite, coarse-grained magnetite, hematite, orthoclase, biotite, and sanidine are generally classed as hypothermal minerals under ordinary conditions. Molybdenite is generally indicative of a fairly high temperature, but is commonly found in intermediate temperature veins, and occurs in at least one epithermal deposit, Cripple Creek, Colorado.[12]

Vanderwilt [18] has classified molybdenite occurrences in the western United States as: (*a*) pegmatite and aplite dikes, (*b*) quartz veins, and (*c*) as an accessory mineral in gold and copper deposits. He states:

> In pegmatite and aplite dikes the molybdenite is generally in radial clusters or flakes up to 2 inches across scattered irregularly through the rock. Large crystals are more characteristic than smaller ones. The molybdenite has been considered both primary and of hydrothermal origin, probably representing a very late phase of pegmatitic activity. Pegmatite dikes in Nevada, and aplite dikes near Romona, California, and in Utah have been described as containing primary molybdenite.

However, the disseminated molybdenite of the aplite dikes may have been introduced by intimately penetrating high-temperature solutions, with the same genetic relationship here proposed for the Ackley City deposit.

The term "complex pegmatite" has been applied to the pegmatites, in which rare minerals have been concentrated by high temperature solutions immediately following solidification of the pegmatite magma. By analogy, "complex aplite" is here proposed as a general term to include the molybdenite-bearing aplites. Apparently, complex aplites are more restricted in mineralogical types than the pegmatites.

Potash metasomatism, characteristic of the Ackley City deposit, is typical of most or all of the principal molybdenite deposits; fluorite is characteristic of many.

The molybdenite-quartz veinlets of Climax, Colorado,[6] contain abundant orthoclase, some fluorite and sericite.

The "Questa" deposit, New Mexico [19] is similar in some respects to the Ackley City deposit; high grade molybdenite veins carry varying amounts of quartz, fine-grained orthoclase, biotite, chlorite, pyrite, fluorite, rhodochrosite, and calcite. The wall-rock alteration consists predominantly of chloritization; the presence of orthoclase and biotite suggests an enrichment of potash, but no chemical analyses are given. However, Vanderwilt states that the chlorite is analogous to the sericite of some molybdenite deposits.

The Utah Copper deposit, Bingham, Utah,[2b] where molybdenite is accessory to copper, shows a marked potash metasomatism in the form of sericite, and a marked reduction in soda.

Schwartz has noted [15] that all available analyses of basic igneous rocks show an increase in potash during hydrothermal alteration, but that the gains in potash balance the losses in the average intermediate and acid igneous rock. Molybdenite mineralization is apparently a type consistently accompanied by an addition of potash, regardless of the original nature of the host rock. The addition is generally in the form of sericite and/or muscovite. Muscovite may predominate over sericite at moderate to great depths and temperatures but only occasionally does this mineral receive specific attention, possibly due to custom, [illegible] ring to all hydrothermal white micas as sericite, even though a great range in grain size of the mica has been noted.

ACKNOWLEDGMENTS.

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GEOLOGY OF THE NIGHTHAWK PENINSULAR GOLD MINE.

A. RODDICK BYERS.

ABSTRACT.

The Nighthawk Peninsular Gold Mine is situated on Nighthawk Lake, ten miles east of Porcupine, Ontario. The ore bodies lie in or near a stock of albite syenite intruding Keewatin lavas. The deposits consist of carbonate zones with disseminated sulphides, cut by auriferous quartz stringers. Bright-green fuchsite is a characteristic mineral. The geological and mineralogical features are discussed and reference is made to the physical and chemical conditions controlling the position of the ore bodies. Comparisons are made with similar deposits occurring in other parts of Ontario and Quebec, in California, and in Western Australia.

INTRODUCTION.

A DESCRIPTION of the more important features connected with the geology of the Nighthawk Peninsular mine is of interest at the present time because of the successful operation of somewhat similar deposits in the Larder Lake and Porcupine areas.

The discovery, in 1906, of gold and quartz veins on Gold Island in Nighthawk lake, lead prospectors to push some ten miles farther west and discover the now famous Porcupine gold camp. The only property so far developed in the Nighthawk lake area is the Nighthawk Peninsular Gold Mine, which is situated at the extreme southwest end of North Peninsula. The property is developed by a shaft to 440 feet depth and by a winze to 630 feet, with 6 levels. Some $566,000 in gold was produced from 1924 to 1926, when falling grade caused suspension of operations until 1933 when the increased price of gold led to an extensive development program.

The only published report that gives any information on the geology of the mine is Hopkins'[1] report on the Nighthawk Lake Area.

[1] Hopkins, P. E.: Ont. Bur. Mines, 33, Pt. 3: 27–36, 1924.

The writer wishes to express his sincere thanks to Drs. J. E. Gill and W. F. James for financial assistance and for the opportunity to make a detailed study of the geology of the Night Hawk Peninsular mine; to Dr. Gill for his advice and guidance during the preparation of this paper, and to Dr. B. S. W. Buffam for his counsel and assistance at the mine. The writer also wishes to acknowledge his indebtedness to Dr. F. F. Osborne for criticism and advice regarding the petrography connected with this paper; and to Dr. R. P. D. Graham for assistance in the laboratory, at McGill University.

ROCK FORMATIONS.

The rocks in the vicinity of the mine consist of basic volcanic flows of Keewatin age, intruded by an irregular stock-like mass of albite syenite and albite-syenite porphyry, the latter being a border phase of the former. Cutting the flows and the syenite are dikes of quartz diabase.

The basic flows are light to dark green in color, fine to medium grained (2–4 mm.), augite andesites. Thin sections of specimens taken from the more massive, less altered phases of the andesite, show phenocrysts of labradorite and pyroxene embedded in a microcrystalline groundmass composed of saussurite, amphibole, epidote, chlorite, carbonate, and secondary quartz, in varying amounts.

The syenite is fine grained (0.2–0.8 mm.), grey to light brown in color, depending on the amount of carbonate alteration, and either equigranular or porphyritic, the two types grading into one another. In thin section, the unaltered syenite is seen to consist mainly of albite with accessory quartz, quartz and albite in micropegmatitic intergrowth, microcline, apatite, biotite, hematite, pyrite and titaniferous magnetite. Gradations from relatively unaltered syenite to a rock composed almost entirely of secondary minerals, may be observed. Ankerite and calcite are the two most common with minor amounts of quartz, chlorite, sericite, tourmaline, and leucoxene. The even or equigranular varieties have an hypidiomorphic or granitoid texture, the albite forming a few euhedral, but generally subhedral and anhedral grains with inter-

locking, sutured outlines. In the porphyritic variety, the phenocrysts are well developed, euhedral, albite crystals. The groundmass of the various porphyritic facies has several textures. The two most common are an allotrimorphic texture developed by irregular interlocking sutured grains of albite; and an hypidiomorphic texture with small irregular laths of feldspar arranged in a divergent or trachyoid manner. Other textures developed are bostonitic and granophyric.

Similar intrusives of albite syenite are found in nearby areas, associated with quartz and feldspar porphyries, syenites, and granites that are considered to be of Algoman age.

The youngest intrusives are a few narrow (1 to 10 feet), black, fine-grained dikes. Thin sections from a dike 8 feet wide, show a fine grained holocrystalline rock with ophitic texture. The essential minerals are augite and plagioclase feldspar ($Ab_{27}An_{63}$). Small amounts of apatite, titaniferous magnetite and biotite are accessory, while varying percentages of carbonate, hornblende, chlorite, saussurite, epidote, leucoxene and albite are secondary.

The age of these dikes is known to be post-Algoman, from intrusive relationships, lack of deformation and alteration as compared with the albite syenite dikes in the same area. Lithologically these basic dikes are similar to the intrusive diabase of Keweenawan age.

STRUCTURE.

The andesite flows strike N. 75° E. and dip 75° S. The syenite forms a small irregular stock with numerous irregular-branching apophyses (Fig. 1) and is 560 to 600 feet long, and 320 to 400 feet wide. From the 625- to the 300-foot level it is nearly vertical, but above the 300-foot level it flattens abruptly towards the west, so that, between the 180- and 300-foot level, it has a pitch of 36° E. This easterly rake is due partly to the original shape of the intrusive and partly to a series of flat easterly dipping thrust faults that intersect the body above the third level.

The complicated faulting in the mine is difficult to interpret because of the lack of markers. Also three or more distinct periods of faulting have occurred. However, three major fault systems

have been recognized, the most important one, No. 1, strikes N. 70° E. and dips 72° S. The faults of this system are all reverse, are pre-ore, and are characterized by heavy gouge seams and much

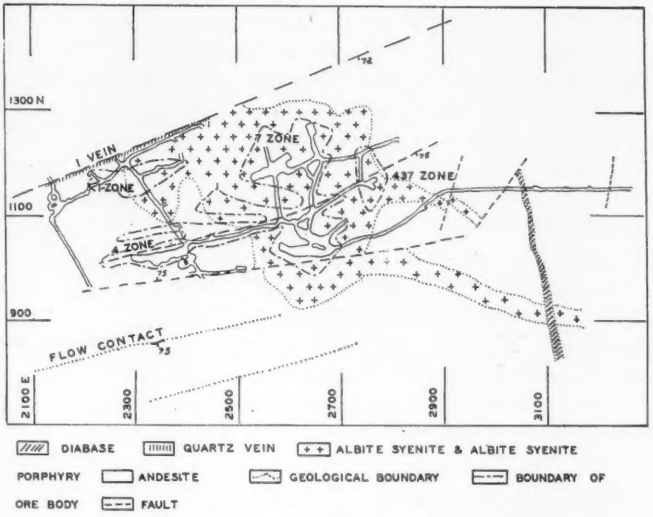


FIG. 1. Geological plan of 425-foot level showing major relationships.

parallel shearing, in some cases, forming sheared zones 20 to 30 feet wide. No. 2 system strikes N. 23° W. and dips 45° E. These are the faults that displace the syenite mass to the west above the 300-foot level. They are reverse faults, with little gouge or shearing, and are post-ore, but are earlier than No. 3 system, which strikes N. 78° E. and dips 75° N. The latter are normal faults with little gouge or shearing. They cut the No. 1 and No. 2 faults and displace the ore bodies.

ORE DEPOSITS.

The ore bodies occur in five zones, designated as No. 1 vein, and Nos. 1, 4, 7, and 437 zones (Fig. 1). With the exception of the No. 1 vein, they are mineralized replacement zones along faults and shears in the andesite and syenite.

The No. 1 vein, from 20 to 40 feet wide and over 600 feet long, consists of a greenish-coloured, sugary quartz. The vein closely follows the hanging-wall of a strong No. 1 fault and thus strikes N. 70° E. and dips 72° S. However, due to the post-ore faulting of the No. 2 and No. 3 systems it has been displaced so that it now has a vertical dip. Gold values are confined to a small shoot where the greenish-coloured quartz is fractured and the fractures filled with a later white quartz, which is slightly mineralized with pyrite and free gold. The shoot pitches 58° E.

The other four ore bodies are carbonatized shear and brecciation zones with disseminated pyrite and arsenopyrite and are cut by numerous irregular quartz veins, which, in places, are numerous enough to form a stockwork. The No. 1 zone is irregular in size and shape, but can be followed from the 180- to the 625-foot level, where it dips vertically and pitches 60° E. The rock of the zone is highly brecciated albite syenite or sheared andesite, both highly altered by carbonatization, sericitization, and silicification. Irregular quartz veinlets cut and cement the brecciated rock. The mineralization is uniform throughout with pyrite, occurring as scattered cubes replacing the rock, and also along the edges of the quartz veinlets. The post-ore faulting is similar to that affecting the No. 1 vein.

The No. 4 ore body lies south of No. 1 zone and joins No. 7 on the east. The body strikes N. 70° E. and the west end rakes 56° E. The average dip between the 300 and 425 levels is 60° S., while between the 425 and 625-foot level it is vertical. The western half lies in sheared andesite and the eastern part is in brecciated syenite. Disseminated sulphides are less abundant than in the No. 1 zone, but fine grained pyrite is common, being associated with quartz veinlets and stringers.

The No. 7 zone is the largest and most important of the ore bodies. Unlike the other zones it lies almost entirely within the syenite. The deposit consists of small anastomosing veinlets of quartz that are so numerous and form such a complete network within the highly carbonatized syenite that the entire zone is a large stockwork. Pyrite is abundant in the wall rock and in the

quartz of the veins. Free gold is visible and occupies minute fractures in the small quartz stringers. The average pitch is 48° SE.

The 437 zone lies just to the east of the No. 7 and the mineralization is identical with that of the larger zone.

Relation of Ore Bodies to Structure.—The ore zones are definitely related to the No. 1 system of faulting with its related shear zones. The No. 1 vein occupies the hanging-wall of a large No. 1 fault. The replacement bodies follow shear or brecciation zones, which are directly connected with strong N. 70° E. faults. Where these faults cut the andesite, the rock is sheared and the mineralizing quartz stringers and veins more or less parallel the shearing. Where they intersect the syenite, the rock is generally brecciated and the quartz veins form an irregular network.

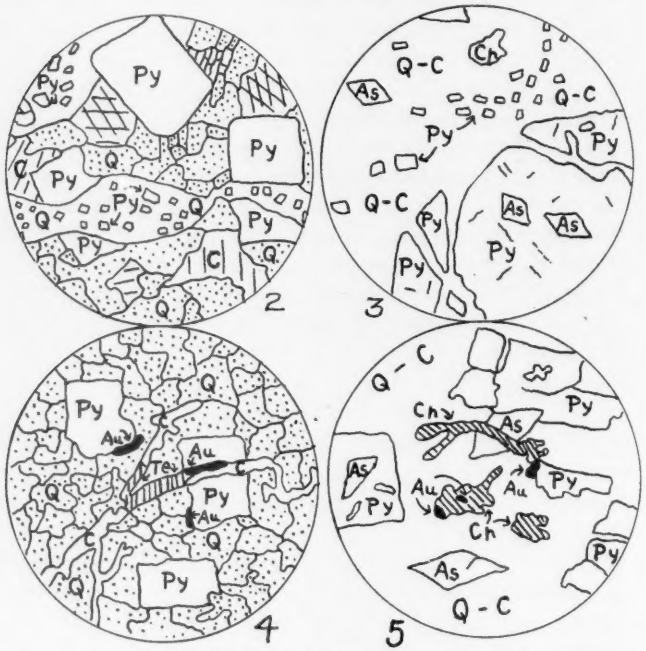
The cause of the easterly rake is unknown, but it may be due to the flat easterly-dipping thrust faults, especially above the 300-foot level. If the latter were pre-ore fractures, then their intersection with the No. 1 system would strike N. 87° E. and pitch 43° E. Hence, this would cause upward-flowing mineralizing solutions to migrate to the west.

Mineralogy.—The mineralogy of the deposit is simple. Pyrite forms 90 per cent of the metallic minerals and arsenopyrite, cobaltite, chalcopyrite, sphalerite, petzite, and gold occur in minor amounts. The gangue minerals are ankerite, quartz and sericite with smaller amounts of fuchsite, chlorite, calcite, albite and tourmaline.

The pyrite occurs in two generations (Fig. 2), the earlier replacing the rock in the form of cubes, which are in places modified in various degrees by the octahedral and pyritohedral forms. The later pyrite is associated with quartz and quartz-carbonate veins and does not occur in such well developed crystals as the first generation. The early pyrite grains vary from 0.001 mm. to 1 mm. in size and average 0.5 mm. On the other hand, the later pyrite averages 0.13 mm. The early pyrite commonly shows signs of deformation, being cracked and cut by later quartz veinlets filled with later pyrite. Arsenopyrite, although rare, is distributed

rather evenly throughout the ore bodies and is closely associated with the earlier pyrite (Fig. 3). Cobaltite is very rare and where observed it was seen to be in close association with the earlier pyrite.

Chalcopyrite and sphalerite occur in minute grains or veinlets closely associated with quartz-carbonate stringers (Fig. 5). Chalcopyrite is next in abundance to pyrite but sphalerite is rela-



FIGS. 2–5. Camera Lucida drawings. Py—pyrite, As—arsenopyrite, Cp—chalcopyrite, Te—petzite, Au—gold, Q—quartz, C—carbonates (ankerite and calite). Magnification 50 ×.

tively rare. Petzite is the only telluride in the ore and is closely associated with free gold (Fig. 4).

Gold was observed in over 25 per cent of the polished sections

examined, which is rather extraordinary in view of the low grade ore. This suggests that all of the gold is free. The gold is in minute grains (.004 mm.), or in small veinlets (.015–.060 mm. wide), and up to .607 mm. in length, replacing or filling small fractures in the pyrite, arsenopyrite, quartz and quartz-carbonate veinlets (Figs. 4, 5).

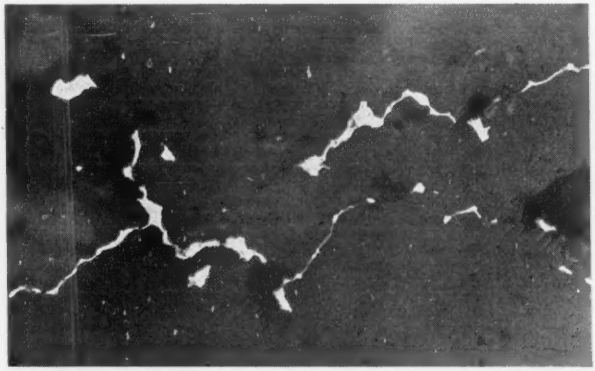


FIG. 6. Microphotograph showing gold in carbonate-quartz veinlet. Magnification 53 ×.

Paragenesis.—The paragenesis, as in most ore bodies, is complex and, in a few cases, textural relationships are difficult to interpret. The deposition of the quartz, carbonates, sericite, fuchsite, chlorite, sulphides, and gold, is thought to have occurred as a more or less continuous process punctuated by periods of fracturing and minor faulting, which formed new entries for the mineralizing solutions. On the basis of this fracturing, the paragenesis may be divided into early, intermediate, and late stages (Fig. 7).

The age relations of the minerals are as follows: During the early stage, quartz followed by ankerite replaced the rocks along fault zones belonging to the No. 1 system. The carbonatization was accompanied by the development of pyrite, arsenopyrite, and cobaltite, and towards the end by sericite, fuchsite, and chlorite. A second stage was commenced by a period of minor faulting and

fracturing. Sericite, fuchsite, and some chlorite continued to be formed during the early part of this stage. The fractures were filled by quartz and minor amounts of ankerite. Associated with

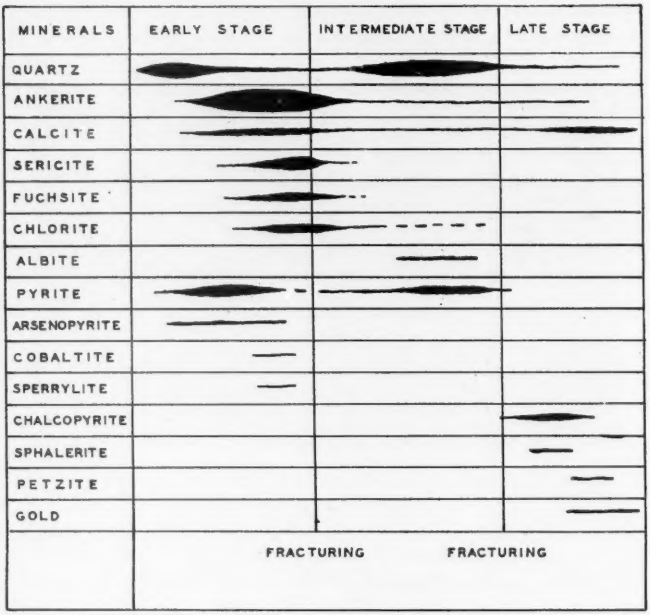


FIG. 7. Paragenesis.

these veins are minor amounts of pyrite and albite. The beginning of the third stage is marked by renewed fracturing, which fissured all previously formed quartz veins and the replaced rocks. These fissures were filled by calcite, a little quartz, and contain small amounts of chalcopyrite, sphalerite, petzite, and gold, which were formed in the order named.

Wall Rock Alteration (*Fig. 8*).—The earliest type of alteration in the sequence of metasomatic changes suffered by the wall rocks during mineralization, was carbonatization accompanied by some

silicification of certain shear and brecciation zones. The carbonatization is the dominant process by which the wall rocks were altered, regardless of whether these rocks were andesite or syenite. It has affected large volumes of rock and is most intensely developed in the ore zones. The carbonate is a pale-brown ankerite and the rocks of the ore bodies have locally been termed "brown carbonate."

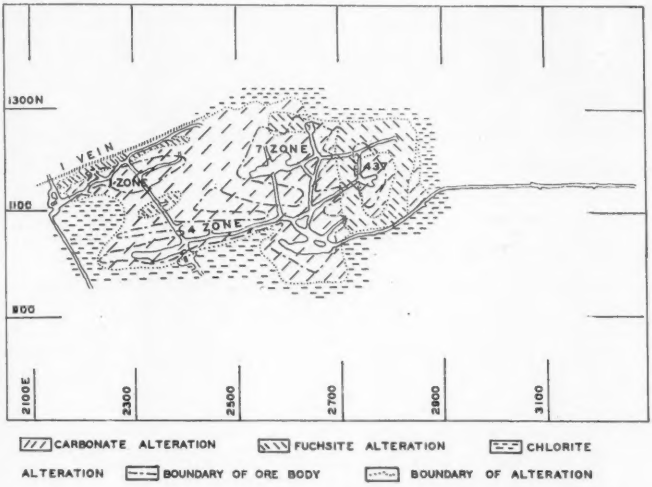


FIG. 8. Plan of 425-foot level showing relation between ore and alteration zones.

Second in importance and more or less accompanying the carbonatization, is the development of sericite and pyrite. The sericite occurs as very fine scales or fibers cutting and replacing the other minerals. Towards the margins of the ore bodies, the sericite becomes less abundant and chlorite takes its place. In certain areas, the sericite grades into fuchsite, a bright-green chromiferous mica, and the rock is then called "green carbonate."

The fuchsite alteration occurred slightly later than the main period of carbonatization. Gold values are lower wherever fuch-

site is found in the ore bodies. This is thought to be due to the leaching by the fuchsite-bearing solutions, of pyrite, the presence of which favored the precipitation of gold.

Chloritization in places is intense and forms a zone of chloritized rock around the margins of the carbonate zones, representing a gradual decrease in the intensity of alteration away from the highly altered ore zones.

The alterations of the wall rocks took place at relatively high temperatures, previous to the deposition of the gold. The latter was introduced during the final stage of mineralization, at much lower temperatures.

Chemical analyses of the fresh and altered syenite, show a notable decrease in silica, soda, and alumina with the exception of the chloritized rock in which case some alumina has been added. This decrease in the above constituents is accompanied by a large increase in ferrous iron, magnesia, lime, potash, and water. Carbon dioxide has been added in large amounts. Cr_2O_3 is not found in the unaltered rock and has been added to the "green carbonate," and to a much lesser extent to the "brown carbonate" and not at all to the chloritized rock. Ferric iron was decreased in the first two types but increased slightly in the chloritized rock. Sulphur is low in the "green carbonate" as would be expected because of the leaching of the pyrite. Increases in the percentage of sulphur have taken place in the other two types, notably in the "brown carbonate." The chemical changes in the major constituents are shown diagrammatically in Fig. 9.

Influence of the Wall Rock on Ore Deposition.—The wall rock in places controlled the position and development of ore-shoots. At the Nighthawk mine there is no change in the average tenor of the ore in passing from syenite into andesite or andesite into syenite. Therefore, it may be concluded that in this case at least, the chemical character of the wall rock, in itself, did not have any effect on the position or grade of the ore shoots.

The physical character of the wall rock had an indirect effect on the distribution and deposition of the ore. Where the major fault zones intersect the rocks, the andesites are generally sheared

and rarely brecciated, but the syenite is commonly brecciated and seldom sheared. Since brecciation favors replacement, the largest carbonate bodies that were formed early in the sequence of min-

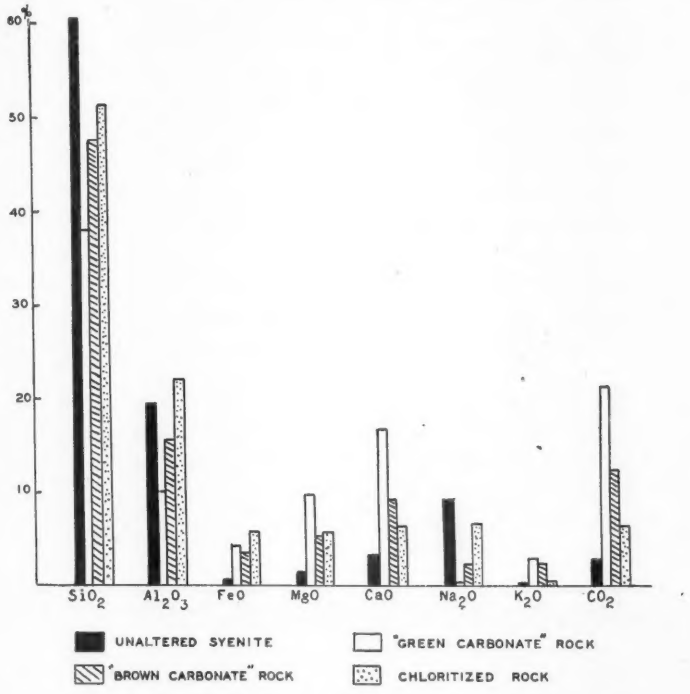


FIG. 9. Diagram of chemical changes in wall rock alteration.

eralization, lie within the syenite. Subsequent movements developed fractures in the carbonatized zones but not in the unaltered rocks, thus permitting selective permeation by the gold-bearing solutions through the carbonitized zones. Hence, the size and shape of the ore bodies largely depended on the size and shape of the carbonate zones, which in turn were dependent on the physical character of the rock that was replaced by the carbonates.

Source and Age of the Mineralizing Solutions.—The general character of the mineralization of the ore bodies indicates that the gold-bearing solutions were derived from a magmatic source. Since the mineralization took place at some time after the solidification of the syenite, it is not directly related to this intrusive, which represents, in all probability, a locus of weakness in an area that connected with an underlying magma reservoir that was the source of both the syenite intrusive and at a later date the ore-bearing solutions.

From the field relations at the mine it is certain that the ore is post-syenite but pre-diabase. The gold mineralization in the Porcupine area lying immediately to the west is considered by most workers to have occurred shortly after the intrusion and solidification of the Algoman intrusives. Since the syenite is genetically related to these Algoman intrusives, it is natural to assume that the ore bodies in the Nighthawk mine are approximately the same age as those in the Porcupine area.

Classification.—There is a close resemblance between this deposit and the "grey ore" bodies of the Mother Lode district, which have been classified by Lindgren [2] as mesothermal. On the other hand, it is also similar to those developed at Larder Lake, which have been described by Lindgren [3] under hypothermal deposits.

In considering only the mineralogy of the ore zones, the diagnostic minerals of the high temperature class are found to be either rare or entirely absent. The association of pyrite, chalcopyrite, arsenopyrite and sphalerite is characteristic of intermediate temperatures.

With regard to metasomatism, the widespread carbonatization and sericitization accompanied by pyritization is typical of intermediate temperatures and pressures. The sericite occurs in extremely small foils and has not been observed in large scales typical of higher temperature formation.

It may be concluded that the deposit belongs in the mesothermal

[2] Lindgren, Waldemar: Mineral Deposits, 3rd Edit., p. 622.

[3] *Op. cit.*, p. 753.

group, that is, mineralization took place at intermediate temperatures and pressures.

SIMILAR OCCURRENCES.

Deposits similar to those of the Nighthawk occur (1) in other localities throughout Northeastern Ontario and Western Quebec,[4] in Canada; (2) in the southern half of the Mother Lode district of California; and (3) in Western Australia. The more important deposits in Canada are found in an area stretching east from Lake Kamiskotia in the district of Cochrane as far as the Kinojevis River in Quebec. Within this area, to date, only the deposits at Larder Lake and Nighthawk Lake have proved to be of economic importance.

In Western Australia, quartz-ankerite-fuchsite bodies are found in (1) Pilbara,[5] (2) Murchison,[6] and (3) Coolgardie[7] Goldfields. They have proved of economic importance in the latter two districts.

The deposits of Australia and Canada are of pre-Cambrian age, and are commonly associated with dikes and small bodies of acidic rocks, rich in soda, which have intruded volcanic greenstones. The deposits may occur within the intrusive, or may lie partly within the intrusive and partly in the greenstone, or they may be found entirely in the intruded basic volcanics.

The deposits of the Mother Lode district,[8] however are of

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[5] Maitland, A. G.: The geology and mineral resources of the Pilbara Goldfield. Geol. Surv. West. Aust., Bull. 24: 40, 1908. Woodward, H. P.: The geology and ore deposits of the West Pilbara Goldfield. Geol. Surv. West. Aust., Bull. 41, 1909.

[6] Clarke, E. deC.: The geology and ore-deposits of Meekatharra, Murchison Goldfield. Geol. Surv. West. Aust., Bull. 68, 1916.

[7] Blachford, T., and Jutson, J. T.: The mining geology of the Kanowna main reef line, Kanowna North-East Coolgardie Goldfield. Geol. Surv. West. Aust., Bull. 47, 1912. Feldtmann, F. R.: The auriferous lodes of the Gibralter district, Coolgardie Goldfield. Geol. Surv. West. Aust., Bull. 91, 1925.

[8] Knopf, Adolph: The mother lode system of California. U. S. Geol. Surv. Prof. Pap. 157, 1929. Ransome, F. L.: Description of the mother lode district, California. U. S. Geol. Surv., Atlas, Mother Lode Folio, No. 63, 1900.

Mesozoic age. Also, they are unlike those of Canada and Australia in that they are not associated with small intrusive bodies but occur in shoots that lie adjacent to the main quartz fissure veins, either in the foot-wall or the hanging-wall. In all other respects they are similar.

In all the deposits, the mineralogy is simple. Native gold and tellurides are the essential constituents of economic importance. Pyrite is the most common metallic mineral, with arsenopyrite next. Chalcopyrite, sphalerite or galena, or both, are always present in small quantities. Among the gangue minerals ankerite and fuchsite (mariposite in the Mother Lode) are the most characteristic.

The ore bodies lie in zones of local shearing and brecciation along which the rocks have largely been replaced by carbonate, sericite, and quartz. The lodes are generally in the form of irregular stockworks with numerous small quartz veins and stringers that carry most of the gold.

SUMMARY.

In the Nighthawk Mine, the ore bodies are found in or near a small stock of albite syenite, which is intrusive into andesite flows that strike N. 75° E. and dip 75° S. The ore zones are closely related to a system of major faults and accompanying shear zones that strike N. 70° E. and dip 75° S.

With the exception of the No. 1 vein, which is a large barren quartz vein mineralized by secondary quartz carrying gold, the ore bodies are carbonatized shear and brecciation zones with disseminated pyrite and arsenopyrite, and are cut by numerous quartz veins, which in places form a stockwork. The ore bodies occur in the andesite and syenite but attain their greatest size in the syenite.

The mineralogy of the deposit is simple. Pyrite forms 90 per cent of the metallic minerals, and arsenopyrite, cobaltite, chalcopyrite, sphalerite, petzite, and gold occur in minor amounts. The gangue minerals are ankerite and quartz with smaller amounts of sericite, fuchsite, and chlorite.

The paragenesis of the ore is as follows: An early development of ankerite, quartz, pyrite and arsenopyrite, and a little later sericite, fuchsite and chlorite. The last minerals to be deposited were chalcopyrite, sphalerite, petzite and gold.

Carbonatization (ankeritization) has been the dominant wall rock alteration within, and chloritization immediately outside, the ore bodies.

Chemical analysis show a decrease in SiO_2, Na_2O, Al_2O_3, and an increase in MgO, CaO, K_2O, FeO, and H_2O.

The chemical nature of the wall rock had no effect on the distribution of the gold values, and the physical character of the rocks only an indirect effect.

The deposit was formed at intermediate temperatures and pressures, and mineralization occurred shortly after the intrusion and solidification of the syenite, which is of Algoman age. These solutions are believed to have originated from the same magma reservoir as the syenite, but outside of this connection the syenite has no genetic relation with the mineralization.

Similar deposits occur in other sections of Ontario, in Quebec,, the Mother Lode district of California, and Western Australia. These deposits, then, represent a type of wide distribution and, in spite of local variations, of remarkable constant characteristics.

GANANOQUE, ONT.,
May 7, 1940.

DISCUSSION AND COMMUNICATIONS

OPENINGS DUE TO MOVEMENT ALONG A FAULT.

Sir: The recent article "Openings Due to Movement Along a Curved or Irregular Fault Plane" by W. H. Newhouse, in the May issue, offers a broad basis for a practical approach to the problem of the efficient exploration of an ore-bearing shear. Having had occasion to apply similar reasoning to that outlined by Mr. Newhouse, I am much interested in his deductions and in his method of approach.

A point not discussed in the article and which may be of some general interest is that of the determination of the probable attitude of ore shoots occurring in areas of low compression, particularly in relationship to their rake. The theories of rock failure under stress demand some regularity in the relationship of the structures produced, one to another; therefore bodies whose occurrence in a structure is controlled by that structure should bear some definite and predictable relationship to it.

To simplify the discussion let it be assumed that an area has been fractured due to non-rotational compression and that there has been only one major period of movement. Two fracture patterns will result; the first, or primary pattern, will be due to the external compressive force, the second, or secondary pattern, will be due to the combined action of the external compressive stress and to that of an internal rotational stress set up by the drag of the fault walls. It is apparent that there must be a difference in the orientations of the two fracture patterns; it is also apparent that since they are both due to the one primary stress they must possess some common plane or axis. Consideration of theoretical strain ellipsoids for the primary strain and for combined primary and secondary strain shows that the intermediate strain axis alone suffers no change of orientation.

The structures associated with the shear itself, then, form a fracture pattern which is controlled in part by the primary stress and has its intermediate strain axis in common with the primary pattern. Since the intermediate strain axis is the line of intersection between the planes of shearing, elongation and those of the tension cracks it follows that the lines of intersection in the secondary pattern are parallel to those in the primary pattern. Areas of low compression and "open spaces" within the shear may be considered as tension cracks conformable with the secondary pattern. Their boundaries will be their lines of intersection with the walls of the shear and their orientation both as to strike and rake, then, will parallel the strike and dip of the intermediate strain axis. Similar reasoning applied to ore shoots controlled within the shear by planes of schistosity or subsidiary shearing leads to the same conclusion. Therefore a determination of the strike and dip of the intermediate strain axis of the fracture pattern of which the ore-bearing shear is a part will indicate angle and direction of the rake of ore shoots contained in the shear. The mutual relationship of the structures of the primary and secondary fracture patterns to the common intermediate strain axis may answer in part the query with which Mr. Newhouse concludes his paper.

Examples exist of instances where early exploration of a mineral prospect would have been facilitated by some advance information on the probable rake of the ore shoots. If it is borne in mind that the conjugate shearing direction, corresponding to the shear under examination, may appear in the vicinity as cross shearing, cross jointing, fracture cleavage or as the attitude of a stringer zone, and that schistose structure occurring near the shear may be a part of the conjugate fracture system, a tentative determination of the strike and dip of the intermediate strain axis may be made. The indicated probable rake of ore shoots may then be tested by the usual exploratory methods. Such a procedure might be employed particularly profitably in cases where a low angle of rake is suspected.

N. D. McKechnie.

Abitibi, Que.,
May 25, 1940.

A THEORY OF MINERAL SEQUENCE IN HYPOGENE ORE DEPOSITS.

Sir: In an article by the above title which appeared in ECONOMIC GEOLOGY, Vol. XXXV, Nos. 3 and 4, M. C. Bandy directs attention to the fact that observed paragenetic sequences do not always conform to his proposed anion to cation weight ratio theory. One of the offending minerals cited is sphalerite, which is often associated with minerals of mesothermal and epithermal classification. In most of these occurrences sphalerite has been determined to be an early mineral with respect to associated minerals, and from this fact it has generally been inferred that sphalerite is a relatively deep-seated mineral or that it was deposited from a distinctly separate generation of mineralization. Although I believe that many geologists have doubted that sphalerite is a relatively deep-seated mineral in iron-copper bearing mesothermal deposits, I am certain that some geologists arbitrarily assume that the relative order of deposition with respect to time reflects the relative order of intensity of deposition. For example, Bandy, on page 560 in the article cited above, states: "If one grants the hypothesis that there is a rough relationship between age and temperature of deposition of minerals, *i.e.*, that the later minerals in a vein are deposited at lower temperatures than the earlier minerals. . . ." Many other instances could be cited, yet I believe this assumption is not always correct.

Although I regard them as useful, all hypothetical investigations in which an attempt has been made to relate the order of mineral deposition to physicochemical conditions of control have failed to conform to all apparent facts. Bandy's anion to cation weight ratio theory appears to be consistent to facts throughout its application to the genesis of pegmatites, magmatic segregations and hypothermal deposits, but according to his words on page 557, "The actual paragenesis of mesothermal deposits shows many disagreements with the 'normal sequence,' in fact, agreements are not more than 50 per cent, probably less."

I believe that Anthon, Schurmann, Wells, Weigel, and others who fathered the study of physicochemical factors as a key to

mineral deposition, were on the right track, but that these investigations were not conclusive because they did not allow for the effects of external variables and because they may have been required to support both apparent as well as real facts. I believe that Bandy's "normal sequence" theory is called upon to accomplish the same impossible task, and it is evident that as long as there is any possibility that our ideas of genesis include any combination of apparent and real facts, no theory may be expected to apply consistently throughout the entire range of mineral deposition.

This comes back to the assumption that late minerals in a vein are deposited at lower temperatures than earlier minerals. If it can be demonstrated that this assumption is incorrect, then it would appear that there has been an attempt to postulate a theory that will consistently account for both facts and fallacies.

Unfortunately I do not know of any means by which it is possible to prove beyond doubt that sphalerite represents a lower order of intensity of deposition than copper-iron minerals. When these two mineral types are found in intimate association, examination under the microscope indicates that the copper-iron minerals, in point of time, are almost always later than sphalerite. Yet, when a great number of deposits containing these two mineral types are considered, it seems significant to note that the preponderance of copper-iron mineral deposition is zonally beneath a sphalerite zone. While it becomes necessary again to resort to an assumption at this point, the inference that deeper zones (geographically) represent progressively higher orders of intensity of deposition, is substantial evidence opposing the assumption based upon simple sequence relationship. I am aware of the fact that many geologists subscribe to both of these assumptions and although they are clearly inconsistent, an explanation of the discrepancies is accounted for by the introduction of extraneous influences. Thus, the mineralization telescopes, a new surge or generation appears, or reversals in temperature trends are postulated to permit agreement with the observed facts. Although I do not deny that any or all of these interruptions from a simple, progressive cycle may obtain, I believe that it is possible to vis-

ualize the operation of a mineralizing cycle that proceeds in an orderly fashion and yet permits many of the conventional discrepancies to fall into a normal order of expectancy.

Obviously, if any assumption is correct, it will be possible to extend its application in agreement with all observed facts. Thus, since there can be little doubt that sphalerite, in terms of sequence, is earlier than chalcopyrite, how can this be possible when it is assumed that the intensity of deposition of chalcopyrite is the greater? This question brings up a point with respect to the visualization of the nature of a mineralizing system in three dimensions. Certainly physicochemical intensity factors are not constant throughout an entire mineralizing system at any given instant and consequently, with respect to time, it is probable that variable intensity factors obtain that permit one type of deposition at a point simultaneously as another type of deposition occurs at another point.

Referring to a specific mineral district, Sales[1] in 1914 described the zonal relationship at Butte. He observed that sphalerite predominated in transitional and peripheral zones with respect to a deep-seated, central, copper-iron zone. This zonal relationship is repeated in nearly all known occurrences in which these two mineral types are present, and I regard this as very strong evidence supporting the assumption that copper-iron mineral deposition is more deep-seated, in terms of physicochemical intensity, than is sphalerite, and thus if zonation reflects deposition intensity, paragenetic sequence may not always be consistent with the intensity order.

In 1933, in a brief discussion of the Butte District,[2] I proposed a means by which it appeared possible to integrate the complex paragenetic relationships noted there. This was based upon the assumption that the minerals deposited in the deep, central zone at Butte were the result of higher physicochemical intensity than those deposited in peripheral areas. It was also assumed that a certain series of minerals could be deposited consecutively in the

[1] Trans. A. I. M. E., Vol. 46.

[2] Copper Resources of the World, XVI. International Geological Congress, pp. 287–305.

central zone simultaneously with the deposition of a different series in the peripheral localities. A study of mineral relationships at or within the transition zone appeared to support these assumptions. Thus, on page 300, I set out graphically the separately determined sequence characteristics of the central and peripheral zones, and by means of observations within the transition zone (Sales' intermediate zone) it was possible to correlate a time relationship between the two established sequences. In summarizing these relationships on page 301, I stated:

> The foregoing observations suggest that quartz and pyrite were being deposited throughout the central, intermediate and peripheral zones simultaneously at the beginning of the mineralizing epoch. Pyrite mineralization terminated abruptly in all places; in the central zone it was followed by enargite and in the peripheral zone it was followed by sphalerite and subordinate amounts of chalcopyrite. Overlap relationships suggest that enargite and sphalerite were being deposited at the same time from the same mineral system, but predominately in their respective zonal positions.

Within the transition zone there is much evidence to suggest that an early deposition of sphalerite was encroached upon by later iron-copper minerals. Displacement of sphalerite was extensive but not complete and residual cores of this earlier mineral within copper minerals produced occasional wide zones of mixed ores. Redeposition of displaced sphalerite occurred outwardly in the system where I anticipate further work will permit at least in part, a distinction between the original and the redeposited sphalerite based upon an interpretation of variations in iron content. Thus a measurable zonal migration is suggested, which I described on page 302, as follows:

> These conditions are best visualized by assuming that the transition between the central zone minerals and the peripheral series first occurred at a position deep in the intermediate zone. As mineral solutions continued to circulate through the vein system for some reason, possibly a rise in temperature, the transition zone moved outward toward the peripheral zone, so that copper minerals of the central zone encroached upon sphalerite, galena and rhodochrosite. All observations indicate that there were no important reversals in the outward migration of this transition zone and that sphalerite, galena and rhodochrosite have only rarely encroached upon the copper mineral series.

From the above comments, it is evident that I agree with Bandy in the recognition of a discrepancy between the paragenetic se-

quence and the " normal sequence " at Butte. However, my purpose in bringing up the discussion is to point out that this discrepancy may be more apparent than real. Actually, it occurs to me that the difficulty lies largely in an incorrect inference in the use of the term, paragenetic sequence. The term applies specifically to relative age, but I think that it is clearly indicated that an implied inference that relative age and relative physicochemical intensity are synonymous, is in many instances wholly incorrect.

Bandy states that it has been impossible for him to establish any agreement between " normal sequences " and paragenetic sequences in mesothermal and those zones of lesser physicochemical intensity. While it is not my purpose at this time to enter into any discussion of genetic theories, I believe that a comprehensive interpretation of genesis may not be emphatically presented until more definite information is available with reference to mineral relationships in low intensity zones. Perhaps observations have been confined too closely to small specimens with too little regard for zonal habit. In conclusion, I might also suggest that it might be well to consider the selection of a new term, such as Bandy uses, " normal sequence," to apply to physicochemical intensity factors, and thereby relieve some of the breadth of assumed application of the term " paragenetic sequence."

L. H. HART.

BUTTE, MONTANA,
July 29, 1940.

REVIEWS*

Australia, A Study of Warm Environments and their Effect on British Settlement. By GRIFFITH TAYLOR. Pages XV and 455; Plates 4; Figs. 142; Map. E. P. Dutton & Co., New York. 1940. Cloth, 6 × 9. Price, $5.00.

This book contains a fund of information concerning Australia that is of interest to geologists, geographers, climatologists and physiographers. Its first part of 126 pages deals with the history of the discovery of the continent, its evolution, geological and physiographic, the character of its soil and a discussion of its desert areas, especially as compared with the arid regions of the United States. The next 86 pages describe in some detail the twenty natural regions into which the author divides the three great structural areas, with reference to their topography, climate and vegetation. The remainder and much larger part of the volume discusses the spread of settlement and the conditions, geological and otherwise, determining the distribution of the continent's water supplies, and its agricultural, pastoral, mining and other industries.

There is one chapter on the geology, climate and industries of New Guinea and New Zealand and brief descriptions of Nauru, Lord Howe, Norfolk, and Macquarie Islands.

The final chapters contain a summary of the conditions affecting tropical settlement in Australia and a brief discussion of the special characteristics of the constituent elements of its complex population.

The book is well written, abundantly illustrated with diagrams, and its text bears the evidence of its authority.

W. S. BAYLEY.

Sedimentary Petrography. 3rd Edit. By H. B. MILNER. Pp. XXII + 666; Figs. 100; Pls. 52. Thos. Murby & Co., London. 1940. Nordeman Publishing Co., N. Y. Price, $10.00.

The great advance made in the study of the composition of sediments and sedimentary rocks during the eleven years that have elapsed since the

* Books noted under *Reviews* and *Books Received* may be ordered through the Economic Geology Bookshop, W. S. Bayley, Urbana, Ill., but orders for official reports and single copies of Journals should be sent directly to their publishers.

appearance of the 2nd edition of this work (see THIS JOURNAL, Vol. 25, p. 107) is probably best appreciated by comparison of this third edition with the preceding one. Not only is the present volume much larger than the earlier edition but it is much more fully illustrated, the number of topics it discusses is much greater, and the applications of the methods described in it are more numerous. In his preface the author states that in the interval between the appearance of the two editions one of the most revolutionary advances on the purely research side of the subject is in the study of the finer materials of sediments, especially clays and soils with X-ray, spectroscopic and base-exchange technique, although these investigations even now must be regarded as only in their infancy "as regards potentialities for unravelling properties of finely divided mineral matter." He also calls attention to the fact that improvements in the technique of investigating sediments in the laboratory are due largely to the demands of industries and to the present vogue for standardizing raw materials.

To take care of the numerous new research phases of the subject and their many economic applications it was found necessary to revise and rewrite much of the 1929 text in order to eliminate obsolete methods of examination of fine mineral grains and replace them with better and more up-to-date methods. Laboratory technique has been expanded into six chapters covering mechanical analysis, mineral concentration, X-ray manipulation, spectrum work, and methods based on fluorescence, microchemistry, and microscopy. Moreover, in addition to the application of the subjects discussed to stratigraphic correlation, other economic applications are briefly described, such as those concerned with building, road construction, problems of water supply, medical treatment and others employed in the asphalt, ceramic and refractories industries.

Especially noteworthy in the new edition is the improvement in the character of the illustrations, especially of the photomicrographs of individual mineral grains and of thin sections of typical rocks. There are 20 plates of microphotographs of mineral grains separated from different types of sediments and 13 plates of thin sections of rocks. The other plates picture heavy mineral residues from oil sands and from various sedimentary rocks, mainly natural rock asphalts.

The two chapters on methods of testing sedimentary rocks and their detailed petrography are very welcome because they collect in a single place the experiences of many investigators on the subject and cite the articles where more complete details may be found. More significant, however, are those on the principles underlying the differentiation and correlation of sediments, illustrated by the results of the study of heavy minerals in the subsurface stratigraphy of certain oil fields in the United

States, Rumania and Poland and in the surface correlation of certain Cretaceous and Tertiary beds in England, for they lead to the problem of the sources of sedimentary materials and suggest the way this problem may be solved. There is perhaps little difference between the lines of discussion with reference to this matter between the 2nd and 3rd editions, but in the latter the treatment is more comprehensive and more philosophical than in the former. One chapter is devoted to the study of soils and another to that of asphalts, synthetic building stones, cements, concrete, fillers, refractories, water supplies, etc., each discussion being accompanied by a well selected bibliography.

The book closes with eight lists of significant minerals arranged according to some important characteristic, a general bibliography containing the titles of about 700 articles and subject and author indexes.

The 3rd edition of Sedimentary Petrography is a compendium rather than a textbook since it meets the needs of all students interested in sedimentary mineral substances from any point of view. It is an authoritative collection of information on the methods of studying sediments.

W. S. BAYLEY.

The Physical Sciences. By E. J. CABLE, R. W. GETCHELL AND W. H. KADESCH. Pp. 754; Figs. 300. Prentice-Hall, New York, 1940. Price, $5.00 (School, $3.75).

This inclusive volume is aimed to help a general reader interpret intelligently the forces of nature, and weigh values, and understand scientific methods of arriving at the truth.

It deals with forces and motion, light, heat, sound, fluids and matter. It includes chapters on the atmosphere, weather, water, electricity and magnetism, and radioactivity. Chemistry and astronomy receive considerable space and six chapters are devoted to phases of geology, particularly rocks, gradational agents, groundwater, diastrophism, earthquakes, volcanism and historical geology.

The volume presents a good survey of the physical sciences.

Commodities in Industry, 1940 Year Book. Pp. 708; Illust. Commodity Research Bureau, 82 Beaver St., New York, 1940. Price, $7.50.

This complete volume covers 75 commodities with statistics and descriptions of occurrence, methods of production, and conversion to finished form. It contains a wealth of information for those who produce, process, or merchandise. It is a narrative and a valuable reference book of permanent value.

BOOKS RECEIVED.

WILLIAM E. BENSON.

WESTERN AUSTRALIA REPORTS.

No. 11. (Sup. to No. 3) **Lalla Rookh Gold Mine, Pilbara Gold-Field.** K. J. FINUCANE. Pp. 3; 1 map.

No. 12. **The Station Peak Mining Centre, Pilbara Gold-Field.** K. J. FINUCANE. Pp. 5; 5 maps.

No. 14. **The North Shaw Mining Centre, Pilbara Gold-Field.** K. J. FINUCANE. Pp. 2; 1 map.

No. 21. **The Salgash—Apex Area, Pilbara Gold-Field.** K. J. FINUCANE. Pp. 3; 2 maps.

QUEENSLAND REPORTS.

No. 12. **The Claudie River Gold and Mineral Field, Portland Roads Dist., Cape York Peninsula.** E. BROADHURST AND E. O. RAYNER. Pp. 16; 5 maps.

No. 30. (Sup. to No. 12) **Same title.** E. O. RAYNER. Pp. 12; 1 map.

Preliminary Maps and Reports of the Canada Geol. Surv. All published at Ottawa in 1939, price 10 cents each. The maps are advance copy, white prints, subject to revision. The following have been received to date:

39–1. **Beaulieu River Area, Northwest Territories.** J. F. HENDERSON. Pp. 16; 1 map.

39–2. **Opawica Lake and Lewis Lake Map-Areas, Abitibi Territory, Quebec.** G. SHAW. Pp. 7; 2 maps.

39–3. **Mackay Lake Area, Saskatchewan.** M. L. KEITH. Pp. 7; 1 map.

39–5. **Snare River Area, Northwest Territories.** C. S. LORD. Pp. 17; 1 map.

39–6. **Quyta Lake and Parts of Fishing Lake and Prosperous Lake Areas, Northwest Territories.** A. W. JOLLIFFE. 3 maps.

39–7. **Cléricy Map-Area, Abitibi and Témiscamingue Counties, Quebec.** W. C. GUSSOW, J. W. AMBROSE, AND H. C. GUNNING. 1 map.

39–8. **Reindeer Lake and Spalding Lake Map-Areas, Saskatchewan.** L. J. WEEKS. Pp. 4; 2 maps.

39–9. **Waconichi Map-Area, Abitibi and Mistassini Territories, Quebec.** G. SHAW. 1 map.

39–10. **Nelson Area, East Half, British Columbia.** H. M. A. RICE. 1 map.

39–12. **La Pause Area, Abitibi and Témiscamingue Counties, Quebec.** J. W. AMBROSE AND H. C. GUNNING. 1 map.

Correlation Index to Aid in Interpreting Crude-Oil Analyses. H. M. SMITH. Pp. 34; figs. 4. U. S. Bur. Mines, Tech. Pap. 610. Washington, 1940. Price 10 cents. *Develops an equation for a system of rating numbers for purposes of comparison.*

Continuity of the Hardyston Formation in the Vicinity of Phillipsburg, New Jersey. J. C. LUDLUM. Pp. 21; figs. 11. N. J. Dept. Conserv. and Devel., Geol. Ser. Bull. 47. Trenton, 1940.

Geology and Metalliferous Deposits at Kootenai Country, Idaho. A. L. ANDERSON. Pp. 69; pls. 2; figs. 18. Idaho Bur. Mines and Geol., Pamphlet 53. Moscow, 1940. Mimeographed. Reconnaissance geol. map 20″ × 16″, scale 1: 126,720, with red overprint showing geology. *General geology, ore deposits (Zn, Pb, Ag), and descriptions of mines.*

Mississippian Border of Eastern Interior Basin. J. M. WELLER AND A. H. SUTTON. Pp. 93; figs. 10. Illinois Geol. Surv., Rpt. of Investig. No. 62. Urbana, 1940. Reprinted from A. A. P. G. Bull. 24; pp. 765–858; 1940. *Brings together 30 years' work, much of it previously unpublished, by many geologists on the petroleum-bearing beds of the Mississippian in more than fifty 15-minute quadrangles in Ill., Ind., Ky., Mo., and Ia.*

Halfway Lake-Beresford Lake Area, Manitoba. C. H. STOCKWELL AND C. S. LORD. Pp. 67; pls. 1; figs. 2; geol. map in color, in three sheets each 14″ × 18″, scale 1: 12,000. Canada Geol. Surv., Mem. 219. Ottawa, 1939. Price, 25 cents. *General geology and descriptions of gold mines.*

La Prospection et les Gisements d'or de demain, R. GOLOUBINOW, and **Itinéraire de Dakar a Douala,** P. LEGOUX. Pp. 18; figs. 2. Bur. d'Études Gèol. et Min. Coloniales, No. 15. Paris, 1940.

Mining Industry of Yukon, 1938. H. S. BOSTOCK. Pp. 18; pls. 3. Canada Geol. Surv., Mem. 220. Ottawa, 1939. Price 10 cents. *Placer gold operations and production.*

Gold and Manganese Deposits of the Lower Barama River, North West District. D. A. BRYN DAVIES. Pp. 32; pls. 6. British Guiana Geol. Surv., Bull. 14. Georgetown, 1940. Price 24 cents.

Transit Traverse in Missouri, Part 3. East-Central Missouri, 1903–37. J. G. STAACK. Pp. 252; 1 pl.; 2 figs. U. S. Geol. Surv., Bull. 916–C. Washington, 1939. Price, 20 cents.

Rumos Novos em Sondagens Profundas. G. DE PAIVA AND I. C. DO AMARAL. Pp. 60. Brazil Div. de Fom. da Prod. Min., Bol. 36. Rio de Janeiro, 1939.

Justificativas para a Locação de um Poco para Petróleo no Recôncavo, Baía. G. DE PAIVA AND I. C. DO AMARAL. Pp. 23; pls. 6. Brazil Div. de Fom. da Prod. Min., Avulso 40. Rio de Janeiro, 1939.

Saude e População da Região do Gurupí (Pará-Maranhão). J. DE A. LIMA. Pp. 16. Brazil Div. de Fom. da Prod. Min., Avulso 42. Rio de Janeiro, 1939.

La Industria Minera en el Perú 1938. J. HOHAGEN. Pp. 351. Cuerpo de Ing. de Minas del Peru, Bol. 124. Lima, 1938. *Mineral production statistics.*

SCIENTIFIC NOTES AND NEWS

The SOCIETY OF ECONOMIC GEOLOGISTS will meet with the AMERICAN INSTITUTE OF MINING AND METALLURGICAL ENGINEERS in New York City on Feb. 17–20, 1941. Joint sessions with the Mining Geology Committee on February 18 and 19 will include one group of papers on strategic minerals and a second group dealing with other phases of economic geology. Sessions with other divisions of the Institute may possibly be arranged. Q. D. Singewald is chairman of the program committee.

The annual dinner of the Society will be held on Tuesday, Feb. 18, at the Mining Club, 33 Broadway. Following the dinner, G. F. Loughlin will give his presidential address on "Comments on the origin and structural control of mineral deposits."

J. B. TYRRELL, geologist, mining engineer and explorer of the Canadian hinterland, was presented with an honorary degree by Queen's University on Oct. 26.

The PETROLEUM DIVISION OF THE A. I. M. E. held fall meetings at Los Angeles on Oct. 17–18 in cooperation with the Southern California Section and at Tulsa, Oklahoma on Oct. 24–26 in cooperation with the Mid-Continent Section.

The INDUSTRIAL MINERALS DIVISION met Nov. 14–16 in joint session with the Illinois Mineral Industries Conference at Urbana-Champaign, sponsored by the Illinois Geological Survey, the Engineering Experiment Station of the University of Illinois and the State Mineral Industries Committee. The Illinois Natural Resources Building was dedicated at this time.

G. W. RUST, recently geologist for Cerro de Pasco Copper Corp., has been appointed supervising engineer for the Mining Section of the Reconstruction Finance Corp., with headquarters in the Federal Reserve Bank building in San Francisco.

The OKLAHOMA MINERAL INDUSTRIES COMMITTEE and the STATE GEOLOGICAL SURVEY sponsored a conference at Norman on Nov. 19.

H. G. RAGGATT of the New South Wales Department of Mines has been appointed assistant geological adviser to the Government of the Commonwealth of Australia.

E. T. MELLOR, widely known economic geologist of South Africa, particularly in connection with the Rand gold deposits, died at Johannesburg on October 10th.

J. J. JAKOSKY has been appointed dean of the School of Engineering and director of the Engineering Experiment Station of the University of Kansas at Lawrence.

H. ST. J. SUMMERS has been appointed to succeed E. W. Skeats in the chair of geology at the University of Melbourne.

INDEX TO VOLUME XXXV

[NOTE.—In this index the titles of principal papers and the headings of departments, as *Discussions* are in italics.]

Acanthite, 567
Aergirite, 485
Africa, South African geology (review), 192
Ahlfeld, F., on tin, 737
  review of book by, 110
Aikinite and silver enrichment at the St. Louis mine, Butte County, Idaho (Anderson), 520
Akin, W. L., on phosphates, 411
Alaska, gold, albite and gold, 707
Alberta, coal, reserves, 671
Albite, 616
Albite and gold (Gallagher), 698
Albite in perthite, 165
Alkalinization, 343
Allen, E. T., on marcasite, 44
  on mercury, 34, 466
  on solutions, 207, 338, 340
  on sulphides, 648
  on volatiles, 282
  on water, 310
Allen, V. T., on fuller's earth, 871
Alling, H. L., on albite, 51
Alteration, albitic rocks and gold deposits, 729
  alkali diminution, 289
  Camp Albion area, Colo., 496
  gneiss, Virginia, 636
  Horseshoe and Sacramento districts, Colo., 817
  mercury mineralization, 140
  Nighthawk Peninsular Gold Mine, Ont., 1004
  Uchi district, Ontario, 392, 397
Alteration in the Uchi gold area, Rock (Bateman), 382
Alteration on the Comstock Lode, Propylitization and related types of (Coats), 1
Ambrose, J. W., on albite, 703
Analyses (see Chemical analyses)
Analysis, reagents used (review), 910
Andersen, H. C., calculation of vapor pressures, 282
Andersen, O., on aplite, 975
  on pegmatites, 953
  on plagioclase, 179
Anderson, A. L., *Aikinite and silver enrichment at the St. Louis mine, Butte County, Idaho,* 520–533
  on mica, 184
Anderson, R. J., *Microscopic features of ore from the Sunshine mine,* 659–667
Ankerite, 85, 399, 613, 814
Ankeritization, 1004
Anorthosite, 575
  Minnesota (review), 117
  Virginia, 633
Antimony, 501
Anton, E. F., on metals, 370
Apatite, 407
Appalachian Plateau (review), 581
Argentite, 531, 567, 813
Arizona, breccia, Miami mine, 93
Arkansas, State Mineral Survey in (Branner), 761
  phosphate deposits, 422
Arrhenius, S., on water, 783
Arsenic, 501
Arsenopyrite, 599, 618, 1001
Asbestos, 488
Asia, gold, albite and gold, 724
Atwood, W. W., review of book by, 581
Auger, P. E., on sulphides, 790
Australia, geography (review), 1019
  gold, albite and gold, 720

Bain, G. W., on gold, 612
Balk, R., review of book by, 580
Bandy, M. C., *A theory of mineral sequence in hypogene ore deposits,* 359–381, 546–570
  discussion of, 1014
Barba, A. A., on silver, 740
Barite, 809, 813
Barr, H. L., on gold, 724
Barrell, J., on biotite, 393
Bartrum, J. A., on feldspars, 723
Bass, C. E., *The Vaucluse gold mine, Orange County, Va.,* 79–91
Bass, N. W., on salt, 848
Bastin, E. S., on "aplites," 728
  on paragenesis, 759

Bastin—*Continued*
on pegmatites, 185
on silver, 562
on textures, 362
Bateman, A. M., on cassiterite, 605
on galena, 667
on gold, 709
on magnetite, 597
on potash, 731
on scheelite, 556
reviews by, 120, 192
Bateman, J. D., *Rock alteration in the Uchi gold area*, 382–404
on biotitization, 401
Bayley, W. S., on Spruce Pine area, N. C., 52
reviews by, 120, 580, 682, 910, 1019
Bechuanaland, geology (review), 193
Becker, G. F., on albite, 708
on chlorite, 3
on cinnabar, 151
on mercury, 21, 470
on ore deposition, 647
Beede, J. W., on "Equus beds," Kan., 840
Behre, C. H., Jr., lead-zinc deposits, Mississippi Valley (review), 787
on Horseshoe and Sacramento districts, Colo., 797
on Leadville ores, Colo., 190
Beljankin, D. S., on cristobalite, 867
Bell, L. V., on albite, 706
on gold, 704
Ben Day peper, 668
Benedict, M., on solutions, 249
Bennett, H. H., review of book by, 683
Benson, W. E., books received, 911, 1022
Benson boiler, 262
Bentonites and fuller's earths, Abundance and significance of cristobalite in (Gruner), 867
Berg, G., on tin, 732
Berkey, C. P., on Shawangunk Mt. area, N. Y., 753
Bibliography, ground water hydraulics, 940
hypogene ore deposits, paragenesis, 568
magnetite in sulphide ores, 606
molybdenite, Rencontre East, Newfoundland, 994
Spruce Pine area, N. C., 185
Uchi gold area, Ontario, 403
Billingsley, P., on gold, 715
on mineralization, 990
Biotite, 988
Birch, —., on viscosity, 314
Bismuth, 520
bornite-klaprotholite, Concepcion del Oro, Mexico, 687
Blake, W. P., on albite, 710
Blanchard, R., on gold, 722
Blatchford, T., on dikes, 721
on gold, 1009
Bloom, M. C., on polymorphous forms, 466
Bloomer, R. O., review of book by, 578
Boatright, W. C., on soils, 543
Boisbaudran, L., on sphalerite, 516
Boley, C. C., with Cady, G. H., *Methods of recording coal data*, 876–882
Bolivia, Structure and mineral zoning of the Pailaviri section, Potosi (Evans), 737
Bonillas, V. S., on pyrite, 597
Books received, 121, 194, 474, 581, 684, 790, 911, 1022
Borax, Bolivia (review), 114
Borings, Kansas, "Equus beds" area, 856
Bornite, 601
Bornite-klaprotholite relations at Concepcion del Oro, Mexico (Krieger), 687
Bournonite, 665
Boutwell, J. M., on molybdenite, 993
Bowditch, S. I., on solutions, 202, 648
Bowen, N. L., on granite, 282
on immiscibility, 329, 776
on reaction principle, 131
on silica, 982
on silicates, 282
on soda, 734
Bradley, W. F., on silica, 870
Bradley, W. S., on metacinnabar, 35
Bragg, W. L., on cristobalite, 872
Brammall, A., on gold, 725
Brandes, W., on sphalerite, 516
Brandt, J. W., on albite, 716
Branner, G. C., *State mineral survey in Arkansas*, 761–771
Branner, J. C., on phosphates, 422
Bray, R. H., on silica, 870
Brazil, mineral resources (review), 120
Breccia, Cuneiform fragments diagnostic of fault (White), 92
Bretz, J. H., review of article by, 581
Bridgman, P. W., on gases, 255, 312
British Columbia, coal, reserves, 672
galena, Slocan district, 666
Broderick, T. M., on mercury, 45
Brown, J. S., on pyrite, 597
review by, 574

Bruce, E. L., on albite, 700
on gold, 382
Bryan, B., Jr., on spodumene, 962
Buddington, A. F., discussion of paper by, 188
on albite, 708
on lamprophyres, 139
on xenothermal deposits, 296, 499
review of book by, 574
Budgeon, N. F., on sphalerite, 512
Bürg, G., on propylitization, 4
Buerger, M. J., on sodium sulphide, 650
on sphalerite, 758
Burbank, W. S., on solutions, 335
Burlew, —., on magma, 286
Burrows, A. G., on gold, 612
Buschendorf, F. v., on gold, 564
Butler, B. S., discussion of paper by, 188
on Alma district, Colo., 797
on chalcopyrite, 554
on iron, 552
on magnetite, 586, 605
on molybdenite, 993
on molybdenum, 731
Butler, R. D., and Singewald, Q. D., *Zonal mineralization and silicification in the Horseshoe and Sacramento districts, Colorado*, 793–838
Butterfield, H. M., on chalcopyrite, 595
Butts, C., review of article by, 581
Byers, A. R., *Geology of the Nighthawk Peninsular Gold Mine*, 996–1011

Cable, E. J., review of book by, 1021
Cadmium, 501
Cady, G. H., and Boley, C. C., *Methods of recording coal data*, 876–882
Cairnes, C. E., on gold, 709
Cairnes, D. D., on arsenopyrite, 601
Calcite, 85, 490, 614, 624
California, gold, albite and gold, 710
Sixteen to One mine, 461
Calkins, F. C., on Coeur d'Alene district, Idaho, 430
on gold, 713
on magnetite, 599
on propylitization, 16
on tourmaline, 276
Cambi, S. C. L., on sphalerite, 509
Campbell, C. M., discussion by, 670
Campbell, I., on propylitization, 6
Campbell, J. M., on tungsten, 731
Campbell, R., review of book by, 580
Campbell, W., on ores, 562
on paragenesis, 361
Canada, coal reserves, 670
gold, albite and gold, 700
Canadian Shield, Genetic relations of gold deposits and igneous rocks in the (Moore), 127
Canadian Shield (review), 581
Carbonates, 297, 527, 814
Carbonatization, 1004
Cassiterite, 952, 959
Catalogue of Virginia maps (review), 578
Chalcedony, 142
Chalcocite, 494, 602, 691
Chalcopyrite, 492, 526, 530, 594, 619, 758, 813, 986, 1002
Chamberlain, R. T., on gases, 272
Charlewood, G. H., on batholiths, 134
on gold, 612
Chemical analyses —
minerals of spodumene pegmatites, N. C., 957
pyroxene, Colorado, 486
rhyolite, South Dakota, 714
rocks, Newfoundland, Rencontre East area, 971
Uchi district, Ontario, 394
sphalerite, 504
Chemistry, mercury sulphide, 37
reagents used in analysis (review), 910
China, geology (review), 682
Chitarow, N. I., on water, 777
Chlorine, 267
Chlorite, 85, 988
Chloritization, 1006
Christy, S. B., on mercury, 25
Chrustschoff, K., on cristobalite, 873
Cinnabar, 22, 906
Clapp, C. H., on magnetite, 598
Clarke, E. de C., on gold, 1009
Classification, igneous rocks, 575
pegmatites, Spruce Pine, N. C., 57
pre-Cambrian, Canadian Shield, 128
Claussen, G. E., on sphalerite, 503
Clay, Arkansas, sampling survey, 768
Cleavage, 440
Cleavage in the recognition of overturned strata, Use of sedimentation features and (Shenon and McConnel), 430
Cline, J. H., on granodiorite, 632
Coal, Canada, reserves, 670
metamorphism (communication), 109
paleobotany (review), 474
Coal data, Methods of recording (Cady and Boley), 876

Coats, R., *Propylitization and related types of alteration on the Comstock Lode*, 1–16
Cobaltite, 1002
Collins, W. H., on gold, 132
Colloids and ore deposition, 212
Colorado, Ore deposits at Camp Albion, Boulder County (Wahlstrom), 477
Zonal mineralization and silicification in the Horseshoe and Sacramento districts (Butler and Singewald), 793
coal, metamorphism (communication), 109
Commodities in industry (review), 1021
Comstock Lode, Propylitization and related types of alteration on the (Coats), 1
Conolly, H. J. C., on contours, 455
Conservation, soils (review), 683
United States (review), 194
Cooke, C. W., review of article by, 581
Cooke, H. C., on dikes, 138
on gold, 1009
on ores, 701
Copper, bornite-klaprotholite, Concepcion del Oro, Mexico, 687
Cornelius, H. P., on minerals, 387
Correlation, rocks and ores, 698
Craig, R. M., review of book by, 580
Crenshaw, J. L., on marcasite, 44
on mercury, 34, 466
on solutions, 338
on sulphides, 648
Creveling, J. G., on Cerro de Potosi, 740
Cristobalite in bentonites and fuller's earths, Abundance and significance of (Gruner), 867
Cross-bedding, 437
Curtiss, L. F., communication by, 571

Daly, R. A., on granite, 282
Dams (review), 471
Dana, E. S., on klaprotholite, 688
on spodumene, 964
Dana, J. D., on spodumene, 964
Dane, —., on viscosity, 314
Darcy, H. P. G., on water, 917
Darcy's Law, 927
Darrah, W. C., review of textbook by, 474
Darton, N. H., on folding, 754
Davidson, S. C., on hiddenite, 949
Day, A. L., on replacement, 243
on solutions, 207, 340
on water, 310
De Ford, R. K., on Graneros formation, 537
De Lapparant, J., on sericite, 173
De Launay, L., on sphalerite, 502
Dendritic textures, 369
De Oliveira, E. P., on albite, 715
Depth in ore deposition (discussion), 188
Derry, D. R., on gold, 704
on replacement, 180
Diagrams—
aids in preparing, 668
Camp Albion ores, Colo., 496
Dupuit formula for water table, 927
gold ores, Nighthawk Peninsular mine, Ont., 1002
magnetite in sulphides, 588, 596
openings along faults, 449
paragenesis, gold, Nighthawk Peninsular mine, Ont., 1004
Horseshoe and Sacramento districts, Colo., 832
Shawangunk Mt. area, N. Y., 758
perthite, Spruce Pine, N. C., 170
phosphate land classification, 427
ripple marks, 435
sphalerite, minor elements, 510
Diamonds, Africa, South (review), 192
Dicke, G., on klaprotholite, 689
Dikes, albite and gold, 701
Canadian Shield, 131
Newfoundland, Rencontre East area, 972
Uchi district, Ontario, 395
Dikes of Amherst County, Virginia, Origin of the nelsonite (Moore), 629
Dimorphous minerals, 566
Directory, Mines Register, 1940 (review), 911
Discussions and communications—
Additional aids in illustration (Ives), 668
Coal metamorphism in the Anthracite-Crested Butte Quadrangles, Colorado (Dapples), 109
Openings due to movement along a fault (Newhouse), McKechnie, 1012
Radioactive standards (Curtiss), 571
Temperature and depth in hypogene ore deposition (Buddington and Butler), Dougherty, 188
The coal reserves of Canada (Campbell), 670

Discussions and communications—Continued
The geochemistry of quicksilver mineralization (Fahey, Fleischer, and Ross), Dreyer, 905
The nature of the ore-forming fluid: A discussion (Graton), Fenner, 883; Ingerson and Morey, 772
Dobell, F. O. S., on gold, 718
Doelter, C., on sphalerite, 501, 513
Dolmage, V., on magnetite, 587
Dolomite, 815
"zebra" structure, 800, 811, 817
Donnay, J. H. D., on chalcopyrite, 595
Dougherty, E. Y., discussion by, 188
on albite, 705
on gold, 612, 700
on ore deposits, 698
Douglas, R. E., on silver, 740
Dreyer, R. M., *The geochemistry of quicksilver mineralization*, 17–48, 140–157
discussion, 905
on mercury, 465, 469
Drilling, Vaucluse gold mine, Va., 91
Drysdale, C. W., on chalcopyrite, 594
Dunes, Kansas, "Equus beds" area, 855
Dunn, L. R., analysis by, 265
Duparc, —., on platinum, 547
Dupuit, J., on ground water, 926
Dupuit formula, 926

Eastman, E. D., on iron, 552
Eckel, E. B., review by, 471
Editorials—
Developments on and around the Witwatersrand (du Toit), 98
Edwards, A. B., on dikes, 722
Elements in sphalerite, Minor (Stoiber), 501
Elias, M. K., on loess, 853
Ellestad, R. B., analysis by, 979, 981
Emmons, S. F., on dolomite, 800
on Horseshoe and Sacramento districts, Colo., 797
on ore deposition, 647
on ore deposits, 698
Emmons, W. H., on Cerro de Potosi, Bolivia, 748
on gold, 713, 715
on mercury, 38
on mineralization, 990
on pipes, 336
on replacement, 906
on solutions, 324
on sphalerite, 369
on tourmaline, 276
Enargite, 566
Engineering and geology (review), 471
Enrichment, gold, Rand district, 107
mercury deposits, 45
St. Louis mine, Idaho, 531
Vaucluse gold mine, Va., 88
Enrichment at the St. Louis mine, Butte County, Idaho, Aikinite and silver (Anderson), 520
Epi-propylites, 16
Epithermal deposits, paragenesis, 564
"Equus Beds" area in south central Kansas, Geology and groundwater resources of the (Lohman and Frye), 839
Erosion, soils, United States (review), 683
Ettlinger, I. A., on enargite, 561
Eutectic textures, 368
Evans, D. L., *Structure and mineral zoning of the Pailaviri section, Potosi, Bolivia*, 737–750
Evans, R. D., communication by, 571
Evans, R. M., on iron, 552
Experiments, lead and zinc sulphides, 648
mercury mineralization, 22, 40

Fahey, J. J., communication by, 465
Farmin, R., on jasperoid, 823
Fault breccia, Cuneiform fragments diagnostic of (White), 92
Fault plane, Openings due to movement along a curved or irregular (Newhouse), 445
Faults, California, Sixteen to One mine, 461
Comstock district, Nev., 12
Horseshoe and Sacramento districts, Colo., 803
Mississippi Valley lead-zinc deposits (review), 786
Nighthawk Peninsular Gold Mine, Ont., 998
openings along a fault, 1012
Raub area, Pahang (review), 680
Sunshine mine, Idaho, 660
Federated Malay States, Raub district, Pahang (review), 675
Feiser, —., on sphalerite, 516
Feldspar, 952, 960
anorthosites, Minnesota (review), 117
composition, Spruce Pine, N. C., 161
Spruce Pine, N. C., 183
Feldtmann, F. R., on gold, 1009
Fenner, C. N., on silica, 255
on textures, 368
on volatiles, 271

Ferguson, H. G., on albite, 711, 712
on cassiterite, 605
on faulting, 456
on gold, 563, 710
on potash, 731
on Sixteen to One mine, Calif., 447, 461
on veins, 447
Ferguson, J. B., on molybdenum, 731
Fermor, L. L., review by, 675
Fersman, A., on pegmatites, 950
Fessler, A. H., on kyanite, 159
on Spruce Pine area, N. C., 52
Feuchere, L., on pyrite, 597
Field geology, (review), 580
Fishel, V. C., on water, 918
Fitzpatrick, E. G., on soils, 543
Fleischer, M. J., communication by, 465
Flint, R. F., review of book by, 473
Florida, phosphate deposits, 411
Fluorine, 267
Fluorite, 491, 988
Folds, cleavage of folded beds, 442
Horseshoe and Sacramento districts, Colo., 803
Folinsbee, R., analyses by, 971, 979, 981
Foreman, F., on iron disulphide, 26
Fracturing, Hollinger veins, Ont., 625
Rencontre East area, Newfoundland, 983
Fragments diagnostic of fault breccia, Cuneiform (White), 92
Fraser, H. J., on mercury, 908
Freeman, H., on sulphides, 647
Frommurze, H. F., on albite, 716, 731
Frye, J. C., with Lohman, S. W., *Geology and ground-water resources of the "Equus Beds" area in south central Kansas*, 839–866
on loess, 852
Fuchsite, 1005
Fuller's earths, Abundance and significance of cristobalite in bentonites and (Gruner), 867

Galena, 492, 526, 599, 619, 646, 665, 757, 812
sheared galena, 666
Gallagher, D., *Albite and gold*, 698–736
books received, 121, 194, 474, 581, 684, 790
on gold, 719
review by, 192
Gallium, 501
Galpin, S. L., on feldspar, 185
Gannett, R. W., on albite, 711
on gold, 563
Gansser, A., review of book by, 473
Gardner, D. L., on albite, 724
Gardner, J., review of article by, 581
Garnet, 58
Gaseous solubilities, 247
Gas-phase action, 230
Gaudin, A. M., on klaprotholite, 689
Geikie, J., review of book by, 580
Geller, A., on sphalerite, 516
Genesis, Camp Albion ores, Colo., 497
feldspar and mica pegmatites, N. C., 179
gold, Nighthawk Peninsular mine, Ont., 1008
ore-forming fluid, 197
Genetic relations of gold deposits and igneous rocks in the Canadian Shield (Moore), 127
Genth, F. A., on Spruce Pine area, N. C., 52
Geochemistry of quicksilver mineralization (Dreyer), 17
discussion of, 905
Geologic maps—
"Equus beds" area Kan., 843
Horseshoe and Sacramento districts, Colo., 796
Lake Albion, Colo., 481
Newfoundland, Rencontre East area, 968
Nighthawk Peninsular mine, Ont., 999
Spruce Pine, N. C., 176
Uchi Lake district, Ontario, 385
Vaucluse gold mine, Va., 82
Virginia, catalogue (review), 578
Geology and engineering (review), 471
Geology and ground-water resources of the "Equus Beds" area in south central Kansas (Lohman and Frye), 839
Geology, anorthosites, Minnesota (review), 117
Bolivia (review), 112
Camp Albion area, Colo., 480
China (review), 682
Coeur d'Alene district, Idaho, 431
Comstock district, Nev., 8
"Equus beds" area, Kansas, 842
Horseshoe and Sacramento districts, Colo., 798
mercury deposits, 19
nelsonite dikes, Virginia, 630
Newfoundland, Rencontre East area, 968
Oklahoma Panhandle, 537

Geology—*Continued*
Raub, Pahang, Malaya (review), 675
St. Louis mine, Idaho, 522
Shawangunk Mt. region, N. Y., 753
South Africa (review), 192
spodumene pegmatites, North Carolina, 943
Spruce Pine district, N. C., 53
structure and field (review), 580
Sunshine silver mine, Idaho, 659
textbooks (reviews), 473, 910
Uchi gold area, Ontario, 383
Vaucluse gold mine, Va., 82
Geology of the Nighthawk Peninsular Gold Mine (Byers), 996
Geophysical prospecting, petroleum (review), 473
Rand gold field, 101
Gerez, J. M., on spodumene, 942
Germanium, 501
Getchell, R. W., review of book by, 1021
Gevers, T. W., on albite, 716, 731
on spodumene, 942
Gibson, R. E., on liquids, 288
on quartz, 209
on solutions, 258
Gieseke, E. W., on spodumene, 960
Gilbert, G., on bornite, 601
on mineral hardness, 371
Gillson, J. L., on orthoclase, 979
Gilluly, J., on jasperoids, 820
on magnetite, 595
on water, 283
Goddard, E. N., on Front Range, Colo., 807
on gold, 713
Gold, Albite and (Gallagher), 698
Africa, South (review), 192
California, Sixteen to One mine, 461
New Mexico, Little Fanney mine, 456
paragenesis, Hollinger veins, Ont., 611, 620
Raub Australian Gold Mine, Pahang (review), 675
Transvaal, Witwatersrand district, 98
Gold area, Rock alteration in the Uchi (Bateman), 382
Gold deposits and igneous rocks in the Canadian Shield, Genetic relations of (Moore), 127
Gold Mine, Geology of the Nighthawk Peninsular (Byers), 996
Orange County, Va., The Vaucluse (Bass), 79
Goldschmidt, V. M., on sphalerite, 501
Goodman, C., communication by, 571
Goranson, E. A., on hiddenite, 949
Goranson, R., on melts, 211
on silicates, 281, 326
on water, 778
Gradient formula, 933
Grange, L. E., on analysis of gas, 265
Granitization, 301
Grating textures, 365
Graton, L. C., *Nature of the ore-forming fluid*, 197–358
discussion, 883
on a sulphide mineral, 688
on alaskite, 727
on gold, 612, 714
on magnetite, 601
on ore deposits, 546
on ore zones, 189
on paragenesis, 759
on purpurite, 958
on solutions, 648, 991
on sphalerite, 502
on spodumene, 944
on textures, 362
on veins, 498
Green, W. F., analysis by, 394, 397
Greenland, geology (review), 581
Greig, J. W., on silica, 260, 780
on silicates, 329
Greisenization, 959
Grim, R. E., on illite, 873
on silica, 870
Ground water, Present status of our knowledge regarding the hydraulics of (Meinzer and Wenzel), 915
Ground water in the Oklahoma Panhandle (Schoff), 534
Ground-water resources of the "Equus Beds" area in south central Kansas, Geology and (Lohman and Frye), 839
Grout, F. F., review of books by, 117, 683
Gruner, J. W., *Abundance and significance of cristobalite in bentonites and fuller's earths*, 867–875
on cristobalite, 868
on magnetite, 595
on sulphides, 647
on textures, 362
Guild, F. N., on pyrite, 665
Guise, C. H., review of book by, 194
Gulf Coast, geology (review), 581
Gunning, H. C., on albite, 703, 709
on gold, 704
Gussow, W. C., on gold, 703

Gustafson, A. F., review of book by, 194
Gustafson, J. K., on albite, 720

Hagen, G., on water, 917
Hale-Shaw, H. S., on flow of liquids, 917
Hall, A. L., on gold, 716
Hall, G. M., on Spruce Pine area, N. C., 52
Halogens and mineralization, 268
Hamilton, W. J., Jr., review of book by, 194
Hannay, J. B., on solubility, 778
Harcourt, G. A., on carbonate, 813
on sphalerite, 502
Hard, H. A., on aquifers, 925
Harker, A., on textures, 61
Hart, L. H., discussion by, 1014
on mineral sequence, 1016
Hartley, W. N., on sphalerite, 503
Harwood, H. F., on gold, 725
Hassler, E. L., on sphalerite, 515
Hawkins, —., on steam, 313
Hawley, J. E., on gold, 732
on granodiorite, 702
on magnetite, 599
Haworth, E., on "Equus beds," Kan., 840
Heim, A., review of book by, 473
Henderson, E. P., on Spruce Pine area, N. C., 52
Henderson, J. F., on albite, 382, 704
on gold, 723
Hershey, O. H., on faults, 660
Hess, F. L., *The spodumene pegmatites of North Carolina*, 942–966
on albite, 180
on biotite, 951
on plagioclase, 61
on spodumene, 962
Hessite, 620
Hewett, D. F., on cinnabar, 47
Heyl, G. R., on gold, 707
Hiddenite, 947
Hillebrand, W. F., on sphalerite, 515
Himalayas, Tibet and Nepal (review), 473
History, lead and zinc, Shawangunk, N. Y., 751
Potosi tin district, Bolivia, 738
Texas oil and gas (review), 790
Hjort, J., on ripple marks, 437
Hogarth, J., on solubility, 778
Hollerith punched card system, 877
Hollinger veins, Paragenesis in the (Keys), 611
Holmes, A., on lead, 127
on mineralization, 990
on uranium minerals, 63
Holmes, S., review of translation by, 910
Hopkins, P. E., on gold, 996
Horton, F. W., on mica, 183
Horwood, H. C., on albite, 706
on magnetite, 603
Howe, E., on magnetite, 597
Hulin, C. D., on gold, 564, 730
on metalization, 137
on veins, 446
Hurst, M. E., on albite, 705
on gold, 612
Hutton, C. O., on amphiboles, 388
Hydraulics of ground water, Present status of our knowledge regarding the (Meinzer and Wenzel), 915
Hypogene ore deposits, A theory of mineral sequence in (Bandy), 359, 546
Hypothermal deposits, paragenesis, 554

Idaho, Aikinite and silver enrichment at the St. Louis mine, Butte County, (Anderson), 520
Coeur d'Alene district, 430
phosphate deposits, 423
silver, Sunshine mine, 659
Iddings, J. P., on albite, 51
Igneous rocks, classification, 575
correlation with ores, 699
petrography (review), 192
Igneous rocks in the Canadian Shield, Genetic relations of gold deposits and (Moore), 127
Ihlseng, Z., on zoning, 760
Illustrations, aids to (communication), 668
Imaizumi, Y., on sphalerite, 509
Inclusions, 364
Ingerson, E., discussion by, 772
on critical phenomena, 650
on ore-forming fluid, 884
on solution, 208, 246, 773, 775
Ingham, A. I., *The zinc and lead deposits of Shawangunk Mountain, New York*, 751–760
Intergrowths, klaprotholite and bornite, 693
Iron, magnetite in sulphide ores, 585
Irving, J. D., on dolomite, 800
on silica, 823
Ives, R. L., communication by, 668
on diagrams, 668
Iwanow, L. A., on water, 777

Jacob, C. E., on water, 926
Jacob, K. D., on phosphates, 410, 420
Jasperoids, 820
Jewell, W. B., on sphalerite, 369
on textures, 362
Johannsen, A., on plagioclase, 50, 179
review of book by, 192
Johnson, J. H., on Alma district, Colo., 797
Johnston, J., on marcasite, 44
Johnston, W. A., on albite, 709
Johnston, —., on solutions, 338
Jointing, Rencontre East area, Newfoundland, 984
Joliffe, A. W., on gold, 382
Jonas, A. I., on granodiorite, 631
on Va. Piedmont, 82
Joubert, G. K., on spodumene, 942
Junner, N. R., on albite, 720
on gold, 721
Juretza, F., on sphalerite, 512
Jutson, J. T., on gold, 1009

Kadesch, W. H., review of book by, 1021
Kansas, Geology and ground-water resources of the "Equus Beds" area in south central (Lohman and Frye), 839
Kato, T., on albitization, 730
on gold, 724
on propylitization, 4
Keenan, —., on steam, 313
Keith, A., on granite, 944
on kyanite, 159
on Spruce Pine area, N. C., 52
Kemp's Handbook of Rocks (review), 683
Kent, L. E., on albite, 716
Kentucky, phosphate deposits, 422
Kerr, F. A., on albite, 708
Kerr, P. F., on albite, 731
Keyes, —., on steam, 313
Keys, M. R., *Paragenesis in the Hollinger veins*, 611–628
on albite, 705
Kindle, E. M., geology of Arctic Archipelago (review), 581
King, H. F., on water, 917
Kistiakowsky, G. B., calculation of vapor pressures, 282
Klaprotholite relations at Concepcion del Oro, Mexico, Bornite- (Krieger), 687
Knight, C. W., on ores, 562
on paragenesis, 361
on silver, 562
Knopf, A., on albite, 708
on biotite, 401
on gold, 563, 711, 1009
on greenstone, 399
review by, 117
review of book by, 473
Knox, J., on mercury, 27
Kovarik, A. F., communication by, 571
Krieger, P., *Bornite-klaprotholite relations at Concepcion del Oro, Mexico*, 687–697
on gold, 715
Kristofferson, O. H., on lead and zinc, 647
Krumbein, W. C., review of book by, 121
Krusch, P., on propylitization, 4
Kyanite, 158

Landes, K. K., on albite, 180
on pegmatites, 733
on plagioclase, 179
Laney, F. B., on klaprotholite, 688
on sphalerite, 369
on textures, 362
Langford, G. B., on albite, 704
Larsen, E. S., on albite, 731
on marcasite, 44
Lausen, C., on chalcopyrite, 563
Lawrence, R. E., data on wells, Wichita, Kan., 864
Lazarevic, M., on propylitization, 4
Lead, Mississippi Valley lead-zinc deposits (review), 786
Lead and zinc sulphides in sodium sulphide solutions, Solution and precipitation of (Smith), 646
Lead deposits of Shawangunk Mountain, New York, The zinc and (Ingham), 751
Leão, J., review of book by, 120
Lee, J. S., review of book by, 682
Legget, R. F., review of book, by, 471
Leggette, R. M., on wells, 940
Legrange, J. M., on albite, 716
Leiding, B., on sulphur, 337
Leith, C. K., on cleavage, 440
Leonhard, A., on sphalerite, 512, 515
Leptothermal deposits, paragenesis, 564
Levin, I., on cristobalite, 867
Lind, S. C., communication by, 571
Lindgren, W., on albite, 710, 722
on alteration, 237
on Cerro de Potosi, 740
on cinnabar, 47
on colloids, 830
on feldspars, 182
on galena, 812

Lindgren, W.—*Continued*
on gold, 1008
on magnetite, 586
on mesothermal zone, 661
on mineralization, 202
on molybdenite, 992
on ore deposits, 698
on propylitization, 2
on pyroxene, 497
on quartz, 759
on replacement, 470, 906
on silica, 823, 827
on solutions, 311
on textures, 362
Lindner, J. L., on pyrite, 595
Lithophysae, 277
Llord y Gamboa, R., on sphalerite, 506
Locke, A., on mineralization, 990
Loess, Kansas, 853
Lohman, S. W., and Frye, J. C., *Geology and ground-water resources of the "Equus Beds" area in south central Kansas*, 839–866
Lombaard, B. V., on albite, 716
Longwell, C. R., review of book by, 473
Lonsdale, J. T., on gold, 80
Lord Howe Island, geography (review), 1019
Loughlin, G. F., on dolomite, 800
on Horseshoe and Sacramento districts, Colo., 797
on Leadville ores, Colo., 190
on pyrite, 597
on silicification, 829
Lovering, T. S., on differentiation, 713
on Front Range, Colo., 798, 807
on solutions, 335
Lugn, A. L., on loess, 853
Luzonite, 813

Macgregor, A. M., on albite, 718
MacLean, A., review of book by, 910
MacPherson, E. O., on albite, 723
Macquarie Island, geography (review), 1019
Magma, 205
Magmatic segregation deposits, paragenesis, 546
Magmatic sequences, Canadian Shield, 131
Magnetite, 494, 986
Magnetite in sulphide ores (Schwartz and Ronbeck), 585
Maitland, A. G., on gold, 1009
Malatesta, L., on sphalerite, 509
Manganese, 508
Mansfield, G. R., *Phosphate deposits of the United States*, 405–429
Map, Arkansas, State Mineral Survey, 766
Bolivia, Potosi area, 738
Camp Albion area, Colorado, 479
Florida, phosphate deposits, 411
Kansas, "Equus beds" area, 840
Oklahoma Panhandle, 535
phosphate deposits, Rocky Mtns., 424
Shawangunk Mt. region, N. Y., 752
South Carolina, phosphate deposits, 409
spodumene pegmatites, North Carolina, 945
Spruce Pine district, N. C., 53
Tennessee, phosphate deposits, 419
Vaucluse gold mine, Va., 81
Witwatersrand system, Transvaal, 100
Marcasite, 43, 148, 524, 566, 603
Mather, W. B., on oligoclase, 705
Mather, —., on lead, 751
Mathewson, E. P., on silver, 740
Matson, G. C., on phosphates, 413
Maufe, H. B., on gold, 717
Maurice, C. S., *The pegmatites of the Spruce Pine district, North Carolina*, 49–78, 158–187
on pegmatites, 733
Mawdsley, J. B., on gold, 137
Mawson, D., on gold, 722
McCallie, S. W., on phosphates, 408
McCally, H., on phosphates, 408
McCaughey, W. J., on kyanite, 159
on Spruce Pine area, N. C., 52
McConnel, R. H., with Shenon, P. J., *Use of sedimentation features and cleavage in the recognition of overturned strata*, 430–444
on erythrite, 661
McKechnie, N. D., communication by, 1012
McKee, E. D., review of book by, 118
McKinstry, H. E., on gold, 612
McKnight, E. T., on jasperoid, 821
on sphalerite, 512
McLaughlin, D. H., on gold, 720
on magnetite, 601
Meigen, W., on sphalerite, 512
Meinzer, O. E., and Wenzel, L. K., *Present status of our knowledge regarding the hydraulics of ground water*, 915–941
on ground water, 536
Mellor, J. W., on mercuric sulphide, 24
Mendelssohn, E., on obligoclase, 715
Mennell, F. P., on gold, 717

Mercury, 501
geochemistry of mineralization, 17
discussion, 905
Mertie, J. B., on albite, 708
Merwin, H. E., on silica, 260, 780
on sulphides, 648
Mesothermal deposits, paragenesis, 557
Metacinnabar, 35
Metallogenesis, gold and albite, 734
Metallogenetic epochs, gold, Canadian Shield, 129
Metamorphism, Adirondacks (review), 578
Uchi gold area, Ontario, 387
Methods of recording coal data (Cady and Boley), 876
Mexico, Bornite-klaprotholite relations at Concepcion del Oro (Krieger), 687
Mica, 69, 963, 987
Spruce Pine, N. C., 183
Microcline, 947
Microscopic features of ore from the Sunshine mine (Anderson), 659
Miller, B. L., on silver, 739
Miller, F. S., on gold, 720
Milner, H. B., review of book by, 1019
Mineral deposition, 306
Mineral sequence in hypogene ore deposits, A theory of (Bandy), 359, 546
Mineral Survey in Arkansas, State (Branner), 761
Mineralization, geochemistry of quicksilver (discussion), 905
ore-forming fluid, 197
Rand gold district, 104
Mineralization and silicification in the Horseshoe and Sacramento districts, Colorado, Zonal (Butler and Singewald), 793
Mineralogy, Camp Albion, Colo., 485
Cerro de Potosi, Bolivia, 749
gold, Hollinger veins, Ont., 613
Nighthawk Peninsular mine, Ont., 1001
Horseshoe and Sacramento districts, Colo., 811
magnetite in sulphide ores, 587
mercury deposits, 19
Mississippi Valley lead-zinc deposits (review), 787
molybdenite deposits, Newfoundland, 985
pegmatites, nelsonite area, Va., 637
Spruce Pine, N. C., 58
Raub gold area, Pahang (review), 680
St. Louis mine, Idaho, 524
Mineralogy—*Continued*
Shawangunk Mt. lead-zinc area, N. Y., 757
silver ore, Sunshine mine, Idaho, 661
Vaucluse gold mine, Va., 85
Minerals, conservation (review), 194
Mines Register, 1940 (review), 911
Miser, H. D., on phosphates, 422
Mississippi delta, sediments (review), 119
Mississippi Valley, geology (review), 581
lead-zinc deposits (review), 786
Moehlman, R. S., on gold 712
Molybdenite, 494, 556
Molybdenite deposits of the Rencontre East area, Newfoundland (White), 967
Montana, phosphate deposits, 423
Moore, C. H., Jr., *Origin of the nelsonite dikes of Amherst County, Virginia*, 629–645
on nelsonite, 639
Moore, E. S., *Genetic relations of gold deposits and igneous rocks in the Canadian Shield*, 127–139
on batholiths, 134
on gold, 705
on silver-cobalt ores, 132
Moore, R. C., Mississippi Valley ore districts (review), 788
on "Equus beds," Kan., 851
Morey, G. W., discussion by, 772
on magma, 223, 286, 973
on ore-forming fluid, 884
on silica, 252, 255
on silicates, 263
on solutions, 208, 773, 774
on water, 282
Morgan, P. G., on feldspars, 723
Moritz, H., on sphalerite, 503
Morton, C. C., on dikes, 722
Movement along a curved or irregular fault plane, Openings due to (Newhouse), 445
Mozambique, geology (review), 193
Mud cracks, 433
Muñoz Reyes, J., review of book by, 110
Murdock, J., on a sulphide mineral, 688
on textures, 362
Muscovite, 58, 69, 962, 987
Mushketov, D., on gold, 725
Muskat, M., on flow of liquids, 916

Nagelschmidt, G., on quartz, 873
Nason, F. L., on rutile, 757
Native mercury, 38

Natural gas, Texas, history (review), 790
Nauru, geography (review), 1019
Nebraska, wells at Grand Island, 929
Nelsonite dikes of Amherst County, Virginia, Origin of the (Moore), 629
Nepal, Himalayas, geology (review), 473
Nevada, Comstock Lode, alteration, 1
mercury, Goldbanks deposit, 141
Newfoundland, The molybdenite deposits of the Rencontre East area (White), 967
gold, albite and gold, 707
New Guinea, geography (review), 1019
Newhouse, W. H., *Openings due to movement along a curved or irregular fault plane*, 445–464
communication on, 1012
on albite, 728
on paragenesis, 372
on sphalerite, 506
review by, 786
Newland, D. H., on lead, 751
New Mexico, Little Fanney mine, 456
New York, The zinc and lead deposits of Shawangunk Mountain (Ingham), 751
New Zealand, geography (review), 1019
gold, albite and gold, 722
Karapiti blowhole, 264
Norfolk Island, geography (review), 1019
Norman, J., on spodumene, 960
North America, central lowlands (review), 581
physiography (review), 581
North Carolina, The pegmatites of the Spruce Pine district (Maurice), 49, 158
The spodumene pegmatites of (Hess), 942
Norton, G. H., on Wellington formation, Kan., 847

Oates, F., on gold, 719
Ödman, O. H., on albite, 725
Oklahoma Panhandle, Ground water in the (Schoff), 534
O'Neill, J. J., on magnetite, 598
Ongley, M., on albite, 723
Ontario, gold, Nighthawk Peninsular mine, 996
Uchi gold area, 382
Opal, 142
Openings due to movement along a curved or irregular fault plane (Newhouse), 445
Ore deposition, discussion, 772
temperature and depth, 188
Ore deposits, A theory of mineral sequence in hypogene (Bandy), 359, 546
Ore deposits at Camp Albion, Boulder County, Colorado (Wahlstrom), 477
Bolivia (review), 110
Brazil (review), 120
Camp Albion, Colo., 483
Cerro de Potosi, Bolivia, 748
Horseshoe and Sacramento districts, Colo., 808
mineral sequence (discussion), 1014
Mississippi Valley lead-zinc deposits (review), 786
Nighthawk Peninsular Gold Mine, Ont., 999
Raub area, Pahang (review), 678
Shawangunk Mt. area, N. Y., 754
Vaucluse gold mine, Va., 85
Ore from the Sunshine mine, Microscopic features of (Anderson), 659
Ore shoots, 445
Ore-forming fluid, Nature of the (Graton), 197
discussion of, 772, 883
Ores, Magnetite in sulphide (Schwartz and Ronbeck), 585
correlation with rocks, 698
St. Louis mine, Idaho, 532
Origin, cristobalite, 873
gold, Rand deposits, 104
jasperoids, 827
klaprotholite, 696
lead-zinc deposits, Mississippi Valley (review), 787
molybdenite, Rencontre East area, Newfoundland, 990
nelsonite, Virginia, 641
phosphate deposits, 409
propylitic alteration, 6, 15
Shawangunk Mt. lead-zinc ores, N. Y., 759
Origin of the nelsonite dikes of Amherst County, Virginia (Moore), 629
Orthoclase, 490, 987
Osborne, F. F., on cobaltite, 563
on trondhjemite, 707
Ott, E., on cristobalite, 867
Oxidation, Vaucluse gold mine, Va., 88

Palache, C., on hiddenite, 949
on sphalerite, 513
Paleobotany, textbook (review), 474
Papish, J., on sphalerite, 512
Paragenesis, aikinite, St. Louis mine, Idaho, 526
Camp Albion ores, Colo., 494
gold, Hollinger veins, Ont., 625
Nighthawk Peninsular mine, Ont., 1003
gold and albitic rocks, 729
Horseshoe and Sacramento districts, Colo., 831
hypogene deposits, 548
lead and zinc sulphides, experiments, 656
Mississippi Valley lead-zinc deposits (review), 786
molybdenite, Rencontre East, Newfoundland, 980
ores, Cerro de Potosi, Bolivia, 749
pegmatites, Spruce Pine, N. C., 159, 162
Raub Australian gold mine, Pahang (review), 680
Shawangunk Mt. area, N. Y., 758
silver-lead ores, Idaho, 663
spodumene pegmatites, North Carolina, 951, 956
Paragenesis in the Hollinger veins (Keys), 611
Park, C. F., Jr., on carbonates, 817
on gold, 80, 727
Partridge, F. C., on albite, 716
on spodumene, 942
Patterns, ore minerals, 363
Pegau, A., on allanite, 641
on microcline, 180
on Va. Piedmont, 82
Pegmatites, 216, 227, 886
Lovingston gneiss area, Va., 637
ore minerals in, 733
paragenesis, 549
Pegmatites of North Carolina, The spodumene (Hess), 942
Pegmatites of the Spruce Pine district, North Carolina (Maurice), 49, 158
Penhale, H. J. D., on tin, 737
Pentlandite, 602
Perthite and albite, Spruce Pine, N. C., 165
Peters, C., on sphalerite, 501
Peterson, —., on klaprotholite, 688
Petrography, Comstock district, Nev., 10
igneous rocks (review), 192
sediments (review), 121, 1019
Petroleum (review), 473
Bolivia (review), 113
Brazil (review), 120
Texas, history (review), 790
Petrov, V. P., on cristobalite, 867
Petterssen, S., review of book by, 911
Pettijohn, F. J., review of book by, 121
Petzite, 620, 1002
Phalen, W. C., on phosphate, 420
Phaup, A. E., on gold, 718, 719
Phillips, W. B., on guano, 407
Phosphate deposits of the United States (Mansfield), 405
Photomicrographs—
albite, Spruce Pine, N. C., 166
"bismuth ore," St. Louis mine, Idaho, 529
cinnabar, 144
feldspar, Spruce Pine district, N. C., 65
gold, Hollinger veins, Ont., 622
klaprotholite, 690
magnetite with sulphides, 600
molybdenite, Rencontre East area, Newfoundland, 980
nelsonite, Virginia, 634
ores, Horseshoe and Sacramento districts, Colo., 810
Sunshine silver mine, Idaho, 662
rocks, Camp Albion, Colo., 487
Uchi district, Ontario, 390
spodumene in albite, N. C., 961
sulphide crystals, 652
Physical sciences (review), 1021
Physiography, North America (review), 581
Piggot, C. S., communication by, 571
Pirsson, L. V., on mud cracks, 433
Plagioclase, 58
Plans—
Little Fanney mine, New Mexico, 457
Sixteen to One mine, Calif., 462
veins, Pailaviri, Potosi, Bolivia, 747
Poiseuille, J., on water, 917
Pond, W. F., on phosphates, 421
Posnjak, E., on Spruce Pine area, N. C., 52
Potosi, Bolivia, Structure and mineral zoning of the Pailaviri section (Evans), 737
Potter, J. A., on iron, 552
Potter, —., on steam, 313
Pratt, J. H., on Spruce Pine area, N. C., 52
Pre-Cambrian, Canadian Shield, 128
Precipitation, mercury sulphide, 37

Precipitation of lead and zinc sulphides in sodium sulphide solutions, Solution and (Smith), 646
Preston, E., on spodumene, 966
Prodinger, W., review of book by, 910
Propylitization and related types of alteration on the Comstock Lode (Coats), 1
Prospecting, nelsonites, Virginia, 644
petroleum (review), 473
phosphate deposits, 412, 415, 426
Purdue, A. H., on phosphates, 422
Purpurite, 958
Pyrite, 43, 85, 106, 492, 524, 566, 595, 616, 758, 811, 986, 1001
Pyrometasomatic deposits, paragenesis, 551
Pyrrhotite, 598, 619, 702

Quartz, 58, 85, 142, 163, 173, 488, 613, 624, 663, 758, 814, 973, 987
Quicksilver mineralization, The geochemistry of (Dreyer), 17
discussion of, 905

Radioactive standards (communication), 571
Ramage, H., on sphalerite, 503
Ramdohr, P., on sphalerite, 511
Ransome, F. L., on Coeur d'Alene district, Idaho, 430
on gold, 1009
on magnetite, 599
on mercury, 152
on molybdenite, 992
on paragenesis, 361
on tetrahedrite, 661
Recording coal data, Methods of (Cady and Boley), 876
Reed, C. M., on dolomite, 800
Reed, J. C., on albite, 708
Reid, J. A., on albite, 706
on gold, 712
Reinecke, L., on gold, 716
Replacement, bornite by chalcocite, 695
mercury deposits, 151
pegmatites, Spruce Pine, N. C., 164
Reviews—
Adirondack igneous rocks and their metamorphism (Buddington), Brown, 574
Australia. A study of warm environments and their effect on British settlement (Taylor), Bayley, 1019
Catalogue of topographic and geologic maps of Virginia (Roberts and Bloomer), 578
Reviews—Continued
Commodities in industry. 1940 Year Book, 1021
Conservation in the United States (Gustafson, Ries, Guise, and Hamilton), 194
Contributions to a knowledge of the lead and zinc deposits of the Mississippi Valley region (Bastin *et al*), Newhouse, 786
Descriptive petrography of the igneous rocks (Johannsen), Gallagher, 192
Die Bodenschätze Boliviens (Ahlfeld and Muñoz Reyes), Singewald, 110
Finding and producing oil (A. P. I.), 473
Geology and engineering (Legget), Eckel, 471
Geology and mineral resources of the neighbourhood of Raub, Pahang, Federated Malay States, with an account of the geology of the Raub Australian Gold Mine (Richardson), Fermor, 675
Geology for the layman (MacLean), 910
Geology of China (Lee), Bayley, 682
Geology of North America, Vol. I (Ruedemann and Balk *et al*), 580
Geology of South Africa (DuToit), Bateman, 102
Geology of the anorthosites of the Minnesota coast of Lake Superior (Grout and Schwartz), Knopf, 117
Kemp's Handbook of Rocks, 6th edition (Grout), 683
Manual of sedimentary petrography (Krumbein and Pettijohn), 121
Mines and minerals in Brazil (Leão), 120
Mines register, vol. 20, 1940, 911
Organic reagents used in quantitative inorganic analysis (Prodinger and Holmes), Bayley, 910
Physical sciences (Cable, Getchell, and Kadesch), 1021
Physiographic provinces of North America (Atwood), 581
Principles of sedimentation (Twenhofel), 193
Recent marine sediments; Symposium by Committee on Sedimentation of National Research Council (Trask), McKee, 118

Reviews—Continued
Sedimentary petrography (Milner), Bayley, 1019
Soil conservation (Bennett), 683
Some memories of a palaeontologist (Scott), Bateman, 120
Structural and field geology (Geikie), Bayley, 580
Texas oil and gas since 1543 (Warner), 790
Textbook of geology. Part I, Physical geology, 2nd edition (Longwell, Knopf, and Flint), 473
Textbook of paleobotany (Darrah), 474
Throne of the Gods (Heim and Gansser), 473
Weather analysis and forecasting (Petterssen), 911
Reynolds, O., on flow of liquids, 917
Rhyolite, analyses, South Dakota, 714
Richards, R. W., on phosphates, 425
Richardson, J. A., on albite, 724
review of book by, 675
Ridgway, J. L., on illustrations, 668
Ries, H., on sphalerite, 512
review of book by, 194
Riley, L. B., on silica, 830
Rimatori, C., on sphalerite, 507
Ringsleben, W. C., on gold, 612
Ripple marks, 435
Roberts, J. K., on Va. Piedmont, 82
review of book by, 578
Robinson, H. S., on gold, 612
Rock alteration in the Uchi gold area (Bateman), 382
Rocks, correlations with ores, 698
igneous rocks, petrography (review), 192
Kemp's Handbook of (review), 683
Rocks in the Canadian Shield, Genetic relations of gold deposits and igneous (Moore), 127
Rogers, A. F., on klaprotholite, 688
on magnetite, 589
on textures, 362
Rogers, G. S., on phosphates, 408, 410
Ronbeck, A. C., with Schwartz, G. M., *Magnetite in sulphide ores*, 585–610
Ross, C. P., communication by, 465
on mercury, 469
Ross, C. S., on albite, 731
on albitization, 727
on feldspar, 734
on magnetite, 598
on mineralization, 202
on Spruce Pine area, N. C., 52
on titanium, 629
Ross, W. O. C., on gold, 716
Rubidium, Spruce Pine, N. C., 958
Ruedemann, R., review of article by, 581
review of book by, 580
Russell, H. L., on ground water, 939
Ryan, C. W., on rutile, 629

Sales, R., on zoning, 1016
Sampling, Arkansas State Mineral Survey, 768
Sampson, E., on chromite, 548
Sandberger, —., on klaprotholite, 688
Sanidine, 988
Schaller, W. T., on albite, 180
on purpurite, 958
Scheelite, 555, 616
Scherer, J. A., on sphalerite, 515
Schmedeman, O. C., on gases, 991
on ore solutions, 201, 294
Schmidt, A., on sphalerite, 512, 515
Schnarschmidt, O., on sphalerite, 512
Schneiderhöhn, H., on propylitization, 5
on sphalerite, 507, 511
Schoewe, W. H., on glacial deposits, 852
Schoff, S. L., *Ground water in the Oklahoma Panhandle*, 534–545
on shales, 537
Schofield, S. G., on pyrite, 597
Schröer, —., on critical phenomena, 258
Schuchert, C., on mud cracks, 433
review of article by, 581
Schuette, C. N., on cinnabar, 145
on mercury, 21
Schürmann, E., on sulphides, 370
Schürmann series, 370
Schwartz, G. M., and Ronbeck, A. C., *Magnetite in sulphide ores*, 585–610
on albite, 400
on orthoclase, 979
on propylitization, 2
on textures, 362, 696
review of bulletin by, 117
Scientific notes and news, 126, 196, 476, 584, 686, 792, 914, 1024
Scott, W. B., review of book by, 120
Seager, G. F., on alaskite, 728
Sections—
Bolivia, Cerro Rico de Potosi, 738
Kansas, "Equus beds" area, 857
Little Fanney mine, New Mexico, 457
Mudsill mine, Colo., 805
nelsonite area, Virginia, 642

Sections—*Continued*
Sunshine mine, Idaho, 431
Vaucluse gold mine, Va., 88
Sedimentation, marine (review), 118
textbook (review), 193
Sedimentation features and cleavage in the recognition of overturned strata, Use of (Shenon and McConnel), 430
Sediments, Kansas, "Equus beds" area, 853
marine (review), 118
permeability coefficient, 538
petrography (reviews), 121, 1019
Rand gold district, 105
textbook (review), 193
Semenenko, N. P., on albite, 714
Sericite, 85, 618
Shenon, P. J., and McConnel, R. H., *Use of sedimentation features and cleavage in the recognition of overturned strata,* 430–444
on erythrite, 661
Shepard, C. U., Jr., on phosphates, 410
Shepherd, E. S., on gases, 272
on silica, 260, 780
Short, M. N., on aikinite, 520, 528
on cinnabar, 149
on enargite, 561
on klaprotholite, 688
Silica, 141, 252, 255
Silicification, 820
Silicification in the Horseshoe and Sacramento districts, Colorado, Zonal mineralization and (Butler and Singewald), 793
Silver, Bolivia (review), 114
Nevada, Comstock Lode, alteration, 1
New Mexico, Little Fanney mine, 456
Sunshine mine, Idaho, 659
Silver enrichment at the St. Louis mine, Butte County, Idaho, Aikinite and (Anderson), 520
Singewald, J. T., Jr., on silver, 739
reviews by, 110, 578
Singewald, Q. D., with Butler, R. D., *Zonal mineralization and silicification in the Horseshoe and Sacramento districts, Colorado,* 793–838
Slavin, M., spectrochemical determinations by, 949
Slichter, C. S., on ground water, 939
Smith, E. A., on phosphates, 407
Smith, F. G., *Solution and precipitation of lead and zinc sulphides in sodium sulphide solutions,* 646–658
Smith, G. O., on phosphates, 428
Smith, H. T. U., on dunes, 855
Smith, J. L., on spodumene, 944
Society of Economic Geologists, 124, 913
Soils, conservation (review), 194, 683
Solberg, —., on steam, 313
Soltsien, P., on sphalerite, 503
Solutions, Solution and precipitation of lead and zinc sulphides in sodium sulphide (Smith), 646
Solutions forming ores, 197
Sosman, R. B., on peridotites, 136
South Carolina, phosphates, 408
South Dakota, gold, albite at Homestake deposit, 713
Southern Rhodesia, geology (review), 193
gold, albite and gold, 717
South-West Africa, geology (review), 193
Spectroscopy, lead-zinc ores, Mississippi Valley (review), 789
Spence, H. S., on muscovite, 183
Spencer, A. C., on albite, 708
on biotite, 401
on pyrite, 597
Spencer, E., on orthoclase, 987
Sphalerite, 369, 492, 525, 566, 598, 619, 646, 757, 811, 952, 986, 1002, 1014
marmatite, 599
Minor elements in (Stoiber), 501
Spodumene pegmatites of North Carolina (Hess), 942
Spruce Pine district, North Carolina, The pegmatites of the (Maurice), 49, 158
Spurr, J. E., on gold, 612, 699, 711, 730
on jasperoid, 821
on magnetite, 599
on ore deposits, 136, 555
on ores, 701
on veins, 446
Standards of radioactivity (communication), 571
Stappenbeck, R., on sphalerite, 507
Stearns, N. D., on water, 919
Stein, W. G. A., on gold, 716
Stephenson, L. W., review of article by, 581
Sterrett, D. B., on feldspar, 184
on granite, 944
on Spruce Pine area, N. C., 52
Stevens, R. E., on biotite, 951
Stevenson, J. S., on albitic rocks, 728
Stillwell, F. L., on dikes, 720
on stannite, 556
on tellurides, 557
Stilson, C. B., on sphalerite, 512

Stockwell, C. H., on albite, 180, 706
on feldspar, 50
on gold, 705
Stoiber, R. E., *Minor elements in sphalerite*, 501–519
composition of sphalerite, 789
Stone, J. B., on lamprophyre, 722
Stovall, J. W., on shales, 537
Strata, Use of sedimentation features and cleavage in the recognition of overturned (Shenon and McConnel), 430
Stratigraphy, "Equus beds" area, Kan., 842
Horseshoe and Sacramento districts, Colo., 798
Mississippi Valley lead-zinc deposits (review), 786
Potosi district, Bolivia, 740
Raub area, Pahang (review), 675
Shawangunk Mt. region, N. Y., 753
Structural geology (review), 580
Structure, Adirondacks (review), 577
Cerro de Potosi, Bolivia, 742
Horseshoe and Sacramento districts, Colo., 803
Mississippi Valley lead-zinc deposits (review), 786
Nighthawk Peninsular mine, Ont., 998
Raub area, Pahang (review), 676
Shawangunk Mt. region, N. Y., 753
Structure and mineral zoning of the Pailaviri section, Potosi, Bolivia (Evans), 737
Stuart-Sharpe, C. F., on patterns, 668
Stuckey, J. L., on kyanite, 159
on Spruce Pine area, N. C., 52
Sulphide ores, Magnetite in (Schwartz and Ronbeck), 585
transport, 299
Sulphides in sodium sulphide solutions, Solution and precipitation of lead and zinc (Smith), 646
Sulphur, Bolivia (review), 114
Sunshine mine, Microscopic features of ore from the (Anderson), 659
Survey in Arkansas, State Mineral (Branner), 761
Swanson, C. O., on cleavage, 442
Synthesis, sulphides, experiments, 646

Tabelman, H. A., on iron, 552
Taber, S., on rutile, 629
Tables—
analyses, minerals of spodumene pegmatites, N. C., 957
Newfoundland, Rencontre East area, 971
Tables—*Continued*
rhyolite, South Dakota, 714
sphalerite, 504, 514
costs, State Mineral Survey in Ark., 763
gold mining costs, Va., 90
magnetite in sulpihde ores, 590
mercury compounds, geochemistry, 28
mineral succession, 376
minerals in granite, England, 725
paragenesis, hypogene deposits, 548
phosphate reserves, Florida, 417
plagioclase, composition, Spruce Pine, N. C., 160
stratigraphy, "Equus beds" area, Kan., 844
Horseshoe and Sacramento districts, Colo., 799
Uchi district, Ontario, 384
Tanganyika, gold, albite and gold, 719
Tarr, W. A., on lead, 760
Taylor, G., review of book by, 1019
Telethermal deposits, paragenesis, 564
Tellurides, 620
Temperature in ore deposition (discussion), 188
Tennantite, 812
Tennessee, phosphate deposits, 418
Tenney, J. B., on pyrite, 597
Terchert, C., geology of Greenland (review), 581
Tertsch, H., on mineral hardness, 372
Terzaghi, R. D., on potash, 983
Tetradymite, 620
Tetrahedrite, 526, 665
Texas, petroleum and gas (review), 790
Textures, klaprotholite, 689
ore minerals, 362
pegmatites, Spruce Pine, N. C., 58
Theis, C. V., on water, 921, 925
Thiem, G., on water, 928
Thiem formula, 928
Thomas, W., on phosphates, 411
Thomson, E., on molybdenite, 551
on silver, 562
on textures, 362
Thomson, J. E., on gold, 383
Thorium, 501
Thornthwaite, C. W., on maps, 668
Tibet, Himalayas, geology (review), 473
Tikonowitch, —., on platinum, 547
Tin, Bolivia (review), 114
North Carolina, spodumene pegmatites, 943
ore-forming solutions, 295
Raub area, Pahang (review), 678

Todd, E. W., on dikes, 136
Tolman, C., on veins, 733
on magnetite, 589
Tolman, C. F., on water, 918
Topographic maps, Virginia, catalogue (review), 578
Tornquist, A., on sphalerite, 502
Tourmaline, 615
Trask, P. D., review of book by, 118
Tröger, W. E., on gold, 725
Tsuboi, S., on plagioclase, 51
Turneaure, F. E., on ground water, 939
Turneaure, F. S., on tin, 295
Turner, H. W., on albite, 710
Twenhofel, W. H., on Dakota group, Kan., 848
on ripple marks, 435
review of book by, 193
Tyler, W. H., on gold, 720
Tyndale-Biscoe, R., on gold, 717
Tyrrell, J. B., on gold, 700

Uchi gold area, Rock alteration in the (Bateman), 382
Uglow, W. L., on albite, 709
on bornite, 553
on cobaltite, 563
on galena, 667
Ultrafilter material, 270
Umpleby, J. B., on faults, 660
on magnetite, 594
on wurtzite, 525
Underground water, permeability coefficient, 538
Union South Africa, geology (review), 192
United States, Phosphate deposits of the (Mansfield), 405
central lowlands (review), 581
conservation (review), 194
soils (review), 683
Unwin, W. C., on flow of liquids, 917
Urbain, G., on sphalerite, 502
Urban, H. M., on Spruce Pine area, N. C., 52
Utah, phosphate deposits, 423

Van Aubel, R., on albite, 720
Vanderwilt, J. W., on carbonates, 817
on molybdenite, 731, 992, 993
Van Eeden, O. R., on albite, 716
Vasileff, —., on cristobalite, 868
Vasilevsky, —., on cristobalite, 868
Vaupell, C. W., on mercury, 23
Veinlets, antecedent, 365
Veins, 445
albite and gold, 702
Veins—*Continued*
Camp Albion area, Colo., 498
Cerro de Potosi, Bolivia, 742
gold, Nighthawk Peninsular mine, Ont., 999
Mississippi Valley lead-zinc areas (review), 787
Paragenesis in the Hollinger (Keys), 611
pegmatite veins, 953
Raub Australian gold mine, Pahang (review), 680
St. Louis mine, Idaho, 523
Shawangunk Mt. area, N. Y., 756
tin, Bolivia (review), 115
Ver Wiebe, W. A., on Wellington formation, Kan., 847
Verhoogen, J., on mercury, 46
on sulphides, 647
Vernadsky, W., on sphalerite, 501
Virginia, Origin of the nelsonite dikes of Amherst County (Moore), 629
The Vaucluse gold mine, Orange County (Bass), 79
maps, catalogue (review), 578
Vogt, J. H. L., on granite, 971
on mica, 951
on platinum, 547
Volynski, F. S., on arsenopyrite, 601
Von Bernewitz, M. W., on gold, 723
von Cotta, B., on veins, 446
von Inkey, B., on propylite, 3
Von Kobell, F., on sphalerite, 507
Von Moos, A., on cristobalite, 867
von Richtofen, F., on propylite, 2
Vugs, 363

Wade, F. B., on gold, 719
Wagner, P. A., on gold, 716
on platinum, 547
on silver-lead ore, 663
Wahlstrom, E. E., *Ore deposits at Camp Albion, Boulder County, Colorado*, 477–500
analysis by, 490
on asbestos, 488
on galena, 492
Waldschmidt, W. A., on galena, 666
Wall rock, mercury, mineralization, 40
Waring, W. G., on sphalerite, 512
Warner, C. A., review of book by, 790
Water table, Kansas, "Equus beds" area, 859
Oklahoma Panhandle, 541
Watson, E. H., on pegmatites, 181
Watson, T. L., on granodiorite, 632
on rutile, 629
Watts, A. S., on Spruce Pine area, N. C., 52

Weather analysis and forecasting (review), 911
Weathering, Spruce Pine district, N. C., 55
Weed, W. H., on chalcopyrite, 595
Weidman, S., on lead, 789
Weigel, O., on solubility, 371
Wells, hydraulics, 926
  Kansas, "Equus beds" area, 856
  Oklahoma Panhandle, 543
Wells, R. C., on solubility, 370
Wenzel, L. K., with Meinzer, O. E., *Present status of our knowledge regarding the hydraulics of ground water*, 915–941
  on Thiem formula, 929
  on wells, 929
White, C. H., *Cuneiform fragments diagnostic of fault breccia*, 92–97
White, D. E., *Molybdenite deposits of the Rencontre East area, Newfoundland*, 967–995
Whitehead, W. L., on silver, 562
Whitlatch, —., on phosphates, 422
Wichita, Kansas, water supply, 863
Willbourn, E. S., on gold, 724
Williams, J. S., on phosphates, 428
  on Phosphoria formation, 423
Williamson, J. H., on Otago district, N. Z., 723
Wilson, M. E., on albite, 704, 731
  on gold, 1009
  review of article by, 581
Winchell, —., on paragonite, 730
Witwatersrand mining developments (editorial), 98
Woodward, H. P., on gold, 1009
Wright, C. W., on chalcopyrite, 594
Wright, J. F., on albite, 706
  on replacement, 180
Wright, L. B., on rhyolite, 714
Wurtzite, 525, 567, 655
Wyoming, phosphate deposits, 423

Xenoliths, 118

Yugoslavia, breccia, Bor mine, 94

Zambrana, J., on rhyolite, 740
"Zebra" structure, 800, 811, 817
Zeolitic alteration, 14
Zies, E. G., on magnetite, 603
Zinc, Mississippi Valley lead-zinc deposits (review), 786
Zinc and lead deposits of Shawangunk Mountain, New York (Ingham), 751
Zirkel, F., on propylite, 2
Zonal mineralization and silicification in the Horseshoe and Sacramento districts, Colorado (Butler and Singewald), 793
Zoning, 277
  Butte, Mont., 1016
  metamorphic zoning, Uchi district, Ont., 392
  Mississippi Valley lead-zinc areas (review), 786
  pegmatites, Spruce Pine, N. C., 173
  Shawangunk Mt. lead-zinc deposit, N. Y., 760
Zoning of the Pailaviri section, Potosi, Bolivia, Structure and mineral (Evans), 737

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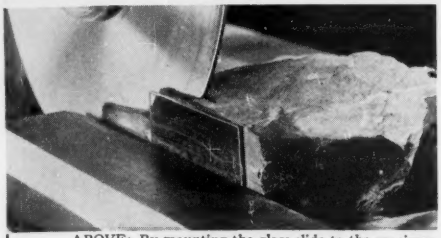

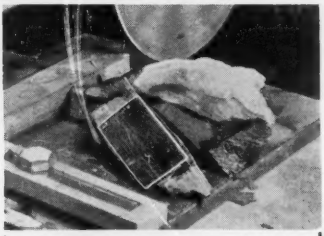

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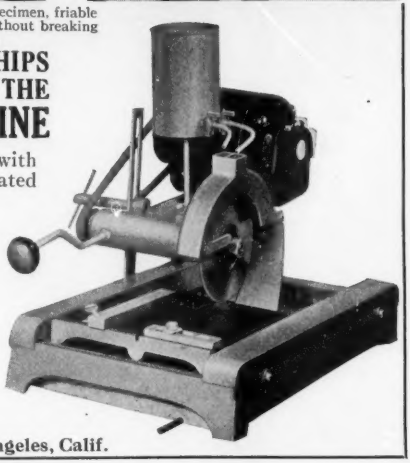

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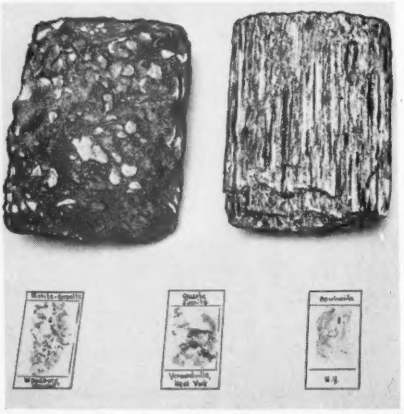

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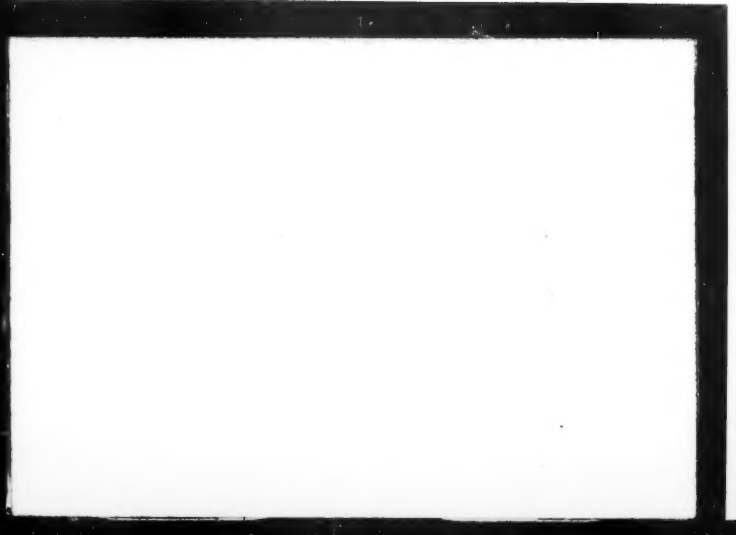
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